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METEOROLOGY

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Division of Physical Sciences
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and
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FOREWORD

It is generally agreed that more attention should be given to research in the middle ground between the Sciences. Geophysics—the study by physical methods of the planet on which we live—is a conspicuous instance of such a middle-ground science, since it shades off imperceptibly in one or another direction into the fields of physics, astronomy, geology, to say nothing of biology, with which the subject of oceanography is closely connected. Some branches of geophysics, such as meteorology, terrestrial magnetism, geodesy and oceanography have long had a more or less independent existence, but it has become increasingly clear that these subjects, and many others, are all parts of geophysics. For various reasons, among which may be mentioned the development of geophysical methods in prospecting for oil and minerals, there has lately been a considerable development of interest in geophysics, but this development has not been matched by the publication in English of systematic treatises on the subject. With these ideas in mind, Dr. J. S. Ames, during his term as Chairman of the Division of Physical Sciences of the National Research Council, was instrumental in organizing in 1926 a large committee to prepare a series of Bulletins on The Physics of the Earth, the purpose being "to give to the reader, presumably a scientist but not a specialist in the subject, an idea of its present status together with a forward-looking summary of its outstanding problems."

In due course sub-committees were formed to prepare reports on the following subjects:

The Figure of the Earth
- Gravity, Deflection of the Vertical and Isostasy
- Tides, Ocean, and Earth
- Variation of Latitude

Seismology
Terrestrial Magnetism
The Age of the Earth
Field Methods for Detecting Unhomogeneities in the Earth's Crust
Internal Constitution of the Earth
Meteorology
Oceanography
Volcanology

That this project, as ambitious as it is important, is now coming to fruition with the publication of these Bulletins is due partly to the skill and farsightedness with which Dr. Ames selected the committee and

assisted in outlining its program; partly to the care and interest with which Dr. Ames' successor, Professor Dayton C. Miller, directed the committee's activities during his term as Chairman of the Division; and particularly to the devotion with which the Chairmen and members of the several sub-committees have carried out their respective assignments. The hearty thanks of the National Research Council and of the readers of these Bulletins is due to the several authors for their efforts.

The volumes will appear serially in the Bulletin Series of the National Research Council, with no particular regard as to sequence, each volume being issued when ready.

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INTRODUCTION: DEVELOPMENT OF THE SCIENCE OF METEOROLOGY

Animal life, and especially that of land animals, is greatly influenced by the weather. This is shown by the hibernation of certain rather low types, such as bears, frogs, and woodchucks, and the seasonal migration of many varieties of birds. Man is no exception to this general sensitiveness to weather conditions. In his early history he selected for his abode regions like the valley of the Nile, or the Tigris and Euphrates, and of the Indus, where extreme cold, destructive winds and drought, were unknown, the latter because the rivers and not the clouds were the local source of the water supply.

As man multiplied on the face of the earth he gradually was forced to take up his abode in regions where it was necessary to protect himself from extremes of cold and from severity of storms, and where he had to depend upon the *former* and the *latter* rain to mature his crops. Probably nothing, except his relations with his brother man, so influenced his life for weal or for woe as did the weather. Of necessity, therefore, he began to observe it, and to make notes in regard to it, so that he might be better able to protect himself and his possessions from harm because of it.

Man soon learned that great power was exerted by the weather elements, a power before which he was helpless. This power he ascribed to gods, good or evil; accordingly, he either worshiped them, or else tried to appease them. Thus, Jupiter Pluvius was worshiped as the rain god, Helios as the sun god, and Boreas as the god of the north wind. The Pueblo Indians of our arid southwest still have a rain dance, by means of which they appeal to their rain god for rainfall.

The ancient Chaldeans and Babylonians made great advances in the study of astronomy. It was natural, therefore, that they should try to connect weather on the earth with the positions of the heavenly bodies, and ascribe to the latter the control of the weather, as also the control of affairs generally on the earth. This was the beginning of astro-meteorology, traces of which are still with us.

There were not wanting, however, those who were able to connect certain phenomena with certain weather types. Atmospherical optical phenomena were among the first so used, and the solar halo as a precursor of rain was early recognized. Likewise, the twilight colors were given a significance, as illustrated in the saying of Jesus to the Sadducees, "When it is evening, ye say, It will be fair weather: for the sky is red. And in the morning, It will be foul weather to-day, for the sky is red and lowering."

There has thus come down to us *weather lore* based upon human experience and intelligent observation, as well as that based upon superstition. To this latter class must be ascribed the long-range forecast that weather conditions of the first twelve days in the year presage the character of the weather for the following twelve months.

The Greeks early applied their philosophy to explaining weather phenomena; and Aristotle's Meteorologia, written at a time when meteorology included meteors, comets, the milky way, earthquakes, and so forth, as related phenomena, was so convincing that it remained the standard work on meteorology for 2,000 years (350 B. C.–1650 A. D.).

At the beginning of the seventeenth century the inductive methods of Francis Bacon, when applied to science, ended the stagnation of centuries, stimulated accurate observational work, and there followed a rapid unfolding of what had hitherto been considered the mysteries of nature. Meteorology shared in this general advancement, as is exemplified by the kite experiments of Benjamin Franklin, and the observations of Thomas Jefferson on Air Temperature and Wind.

At this time observational meteorology was still in its *diary* or *daily journal stage.* Perhaps the earliest of such daily records is found in the *parapegmata,* which evidently dated back to the fifth century, B. C. They were, in effect, weather bulletins, posted in public places, and in addition to summarizing observations, especially of the wind, they also contained predictions, of which the following are quoted from Hellman's The Dawn of Meteorology: *

> Sept. 5. Rising of Arcturus. South wind, rain and thunder.
> Sept. 12. The weather will likely change.

A much more complete daily journal was kept by William Merle, a Fellow at Oxford, between the years 1337 and 1344, A. D. Hellmann concludes that the real object of the journal was to verify weather forecasts which Merle and his colleagues made, and which were based on astro-meteorology, of which they were students.

It was after the invention of the thermometer and the barometer, early in the seventeenth century, that accurate and systematic measurements of the weather elements became possible, although the ancient Greeks had wind vanes much like those now in use. In Chapter II, under the *Instrumental Period,* it is shown how little the value of weather records was appreciated; for at the beginning of the nineteenth century a serious effort to obtain such records was being made at only twelve points in Europe, and ten in North America.

* Quart. J. Roy. Meteorol. Soc., 34: 224 (1908).

In the eighteenth century Halley and Hadley attempted a physical explanation of atmospheric phenomena; and in the nineteenth century Wm. Ferrell, a self-educated mathematician, brought out his great work on Modern Meteorology. Thus was inaugurated the attempt to develop meteorology as a branch of physics, a work that is still going on.

A great impetus was given to this movement by the invention of the electric telegraph, and its utilization in the rapid collection of weather reports. This made possible the construction of daily weather maps, which was undertaken in Prague, in London and in Paris about the middle of the nineteenth century, the object being to forecast the weather. At about the same time the Smithsonian Institution at Washington (1855) commenced the collecting and charting of weather observations. Forecasting from weather charts began in Holland in 1860, and Professor Abbe made the first forecasts in the United States at Cincinnati in 1869.

The methods first employed in forecasting were largely empirical. Certain types of weather maps were observed to be accompanied by certain types of weather, which advanced in a certain direction at a determined average rate. Examples of weather types for the United States have been given by Bowie and Weightman in Supplements to the Monthly Weather Review, Numbers 1 and 4.

Although reasonably successful, the method failed when unusual conditions arose. Gradually a more complete analysis of the forces operating to develop and maintain storms has been developed with gratifying results, as is shown in Chapter VI, The Physical Basis of Weather Forecasting. Much has been contributed to this development in the present generation by such men as Sir Napier Shaw, V. and J. Bjerknes, the late F. M. Exner, and others.

There has been a gradual widening of the field of intensive meteorological observation, such as the careful identification of different cloud types, the altitude at which they usually are formed, and their relation to the passage of areas of high and low atmospheric pressure. The upper layers of air have been explored by means of kites and balloons to determine their temperature, vapor content, and movements. The spectroscope has enabled the atmosphere to be examined to altitudes not attainable by balloons, and its composition, especially as regards the ozone content, has engaged the attention of several able investigators, as is set forth in Chapter I.

The spectroscope, in connection with certain heat measuring devices, has also been used to examine the character and determine the intensity of solar radiant energy as it reaches our atmosphere, and its depletion in its passage through the atmosphere to the surface of the earth. This enables a determination to be made of the energy available to keep in

motion the great atmospheric engine that produces the weather, and also to maintain vegetable and animal life. Details of solar radiation and its rôle are given in Chapter III.

In the early years of the present century man has learned to navigate the air. Sailors on the sea have long profitted from careful attention to weather forecasts, and especially from forecasts of visibility, and of the direction and force of the wind. The pilots of airplanes are even more dependent upon accurate foreknowledge of weather conditions to be encountered on the airways than are their brother navigators upon forecasts of weather at sea. Therefore it is no longer enough to observe atmospheric conditions at the surface of the earth. They must be known to the height at which planes fly. Chapter IV gives an account of the methods by which these upper air data are obtained and utilized.

The requirements of commerce and industry call for the collection of meteorological data in such quantities that its assimilation by scientific investigators has not kept pace with its production. Especially is its use in scientific research retarded by the fact that no international units of measure have been adopted by the different official weather services. In consequence, before any extended investigation can be undertaken, meteorological data, such as atmospheric pressure, temperature, humidity, and wind velocity, as tabulated in different countries, must be reduced to a common unit of measure, at a great expenditure of time and labor. The meteorological section of the International Geodetic and Geophysical Union is endeavoring to bring about the adoption of international units by the meteorological services of the world.

From the foregoing it appears that the science of meteorology is developing in a normal way. Observation has shown that certain events, such as high winds, or rainfall, occur in correlation with certain atmospheric conditions. The law controlling the occurrence of these events is not easily determined, or, in the present state of our knowledge, its application to forecasts of the weather is so cumbersome as to render its use impracticable. Shall we therefore conclude that the theoretical side of meteorology is of academic interest only, and that the practical meteorologist need not concern himself about it? Rather should not the practical meteorologist cooperate with the theorist in his studies by pointing out what is helpful and what requires further development before it can be applied in the solution of his problems? The practical meteorologist should also be ready to supply additional and more accurate data, and in the form required by the research student, to the end that meteorology may be firmly established as a branch of physics, and be worthy of study for its cultural value as well as because of its practical applications.

Chapter V, Dynamic Meteorology, sets forth some of the fundamental physical laws underlying the science of meteorology. It will be noted that certain of these laws are already employed in weather forecasting (see Chapter VI). Others require development before they can be applied to the solution of problems arising almost daily in connection with the work of a meteorological office.

Besides the paper by Hellmann, mentioned previously, I have also been greatly assisted in writing this Introduction by Chapter 1, Volume 1, Manual of Meteorology, by Sir Napier Shaw, and by a paper by Edgar W. Woolard, on The Development of Meteorology as Illustrated by the Rôle of Mathematics in the Progress of Science (Monthly Weather Review for December, 1923, 51 : 645-649).

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CHAPTER I

THE ATMOSPHERE: ORIGIN AND COMPOSITION

WILLIAM J. HUMPHREYS

The most primitive man—the wildest savage—recognizes a wind when out in it, but it took a Greek philosopher to tell us what it is: air in motion. What motion is we know, at least well enough for practical purposes, but what is air? What is that invisible and odorless something we breathe and therefore call atmosphere, the thing that affords us all our weather perceptions and whose states and conditions we have learned to measure and even to foretell? And what was its origin?

ORIGIN OF THE ATMOSPHERE

To start as nearly as possible at the beginning, how and when did the earth ever get an atmosphere—its gaseous envelope? So far as we know there are two, and only two, basic substances, the electron, a certain extremely minute quantity of negative electricity, perhaps plus something else; and the proton, an equally small quantity of positive electricity, also perhaps, plus something else. Of the origin of these entities no one has the slightest idea. They can exist separately, in which case they are electrically very active; or variously grouped together in equal numbers, with increase of inertia but almost total loss of electric force, at least on things external. Every such group that is stable, or even measurably durable, is a chemical element. There are no other elements, and this number is limited. Furthermore, though occurring in unequal quantities, all these elements appear to be distributed throughout the universe. Most of them have been found in the sun, for instance, and we believe the others are there too, even if in such relatively small amounts as to be difficult of detection. Hence, when that other star, some three billion four hundred million years ago, passed so near (within a thousand million miles, perhaps) our sun as to drag off from it by tidal action the masses that coalesced into the planets there were present, and came off together, all the possible elements. That is, the primordial material of the earth, as it was pulled off from, or out of, the sun, consisted of all the elements that now make it up, so that at the very beginning of the independent existence of the earth it had, if not an actual atmosphere, at least the making of one.

It is believed that as the earth mass drew together in a molten sphere the heavier and more refractory substances formed mainly the inner core,

and that the lighter and more volatile elements and compounds formed the rocky shell and the gaseous envelope. In this way, it is believed, the earth pretty soon had oceans and an atmosphere of some kind. However, even if all the gases, hydrogen, oxygen, nitrogen, and others, that can combine with various elements and compounds and form solids had then so combined, leaving neither water nor air, it is certain that before long the earth would have begun accumulating both. It is doing so now through every volcano, every fumerole, and every bubbling spring, and must always have done so since before even the first crust began to form. But it seems unlikely that our present atmosphere actually did come entirely from the molten interior, whether by way of volcanic activity or otherwise, because volcanoes do not, so far as we know, give off free oxygen—volcanic gases captured before there has been any chance for admixture with the air show no trace of it. Furthermore, this element could not exist uncombined in the presence of hydrogen and sulphur at high temperatures, both of which are abundant in volcanic vapors. But we have oxygen; where did it come from? It is known that highly developed green plants produce, under the stimulus of light, a great deal of oxygen from carbon dioxide, a gas abundantly emitted by volcanoes. But most primitive plants do not; they consume it; hence this promising source of free oxygen appears, on close examination, to be most uncertain. However, some forms of lower life thrive in the absence of free oxygen and yet, in the presence of light, evolve it from certain of its compounds, especially water. Again, lightning, which must have occurred from the beginning, frees a little oxygen from water; and so also may ultra violet light. There have been, then, continuously active means of obtaining free oxygen from its compounds since the beginning of the world. Nevertheless, it seems most likely that at the beginning there was more oxygen present than was necessary to use up the free hydrogen and to combine with the available surface materials—enough to do all this and to have a goodly amount left over as free oxygen of the atmosphere. Presumably, too, there were present other primitive gases, especially nitrogen, argon, carbon dioxide and water vapor. But whatever the primitive state of the atmosphere when the earth was first formed, it may be regarded as practically certain that during the whole of the three billion four hundred million years since that time it has been continuously depleted by combination with many things in and of the crust, and also as continuously repleted by their decomposition. It is always changing, but except in respect to water vapor, the change is so small in comparison with the whole that we are not ordinarily aware of it.

It has been argued that the earth could not have retained an atmosphere when molten, or even when dull red. But this is true of only the lightest two gases, hydrogen and helium, and of them only if no other gases were

present in large amounts. A deep atmosphere of water vapor, for instance, would catch any light gas that might leave the earth beneath with an escaping velocity. It could escape only if the outer portions of the atmosphere also were quite hot. At any rate, as soon as a crust, however thin, formed over the earth the supply of heat from beneath was so reduced (the crust being a good insulator) that the upper air necessarily became cool enough to retain the lightest gases. Also, water must soon have begun gathering on the surface. Even if, up to this stage, all helium and free hydrogen had been driven wholly away from the earth, a condition that seems unlikely, there has been since then, and still is, abundant opportunity to accumulate both of them—hydrogen from volcanoes, and helium from radioactive materials everywhere.

Presumably, therefore, the atmosphere is primitive in part—pulled off from the sun with all the other elements—and in part, at least, certainly regenerated inasmuch as every volcano is an active air factory.

COMPOSITION OF THE ATMOSPHERE

Apparent simplicity.—Presumably every one usually thinks of air as being simply air, a homogeneous and single thing. Many of us always think of it that way, as the ancients did, when we think of it at all. Indeed, so far as most of its physical properties and behavior are concerned it shows no obvious complexity. It is simply air in motion that is responsible for a thousand familiar things from the stir of a leaf to the wreck of a house; and simply air that floats the balloon and sustains the aeroplane. In all these matters it usually is quite satisfactory to regard the air as the single substance it ordinarily seems to be.

Evidence of complexity.—This apparent oneness of air does not, however, extend to all physical processes. When we try to liquefy it, for instance, evidences of its complexity soon become amazingly conspicuous. If untreated air is forced through the cooling coils they quickly become choked with ice; and they still clog up when even the dryest air is used, if nothing but the water vapor has been removed—this time with solid carbon dioxide. Then, too, the liquid air itself shows abundant evidences that it is a mixture and not a simple substance like water. What then are the known constituents of the atmosphere, and how and when were they discovered?

Discovery of water vapor.—The earliest considerations of the composition of the air that have come down to us are those of the Greeks. In their speculative philosophy on the composition of objects, they considered air to be one of the four elements (fire, earth and water being the other three) that, singly or variously combined, make up all substances. From this

it might seem that these Greek philosophers regarded the atmosphere as strictly a single thing—the "element" air. Yet it seems that by "air" they meant anything gaseous, or at least that component of it that makes it gaseous, and not necessarily the atmosphere. At any rate, Aristotle,[1] 250 years B. C., says very distinctly, in his work on meteorology, that cloud and rain are caused by condensation from the atmosphere of water vapor that had got there by the evaporation of water at the surface of the earth. He thus makes it very clear that the air consists of at least two things, and that water vapor is one of them.

Water vapor, then, was the first constituent of the atmosphere to be explicitly recognized. Aristotle mentions it, but it is not certain that this discovery was original with him. However, his is the earliest record we have of it, and for that reason, there being no evidence to the contrary, we regard him as one of the discoverers—the earliest one—of the constituents of the air.

Delay of further discoveries.—For more than twenty-two centuries, therefore, and perhaps for much longer, it has been known that the air we breathe consists of at least two things, water vapor and whatever is left after the water is removed. And for more than two thousand years after the days of Aristotle that is all that was known about its composition. Indeed it was practically impossible to push our knowledge of the atmosphere any farther without something of the facilities and methods of the modern laboratory, nor before there had been acquired—very slowly and tediously it was—a fair concept of chemical elements and pure substances. Not until the beginning, then, of the eighteenth century was it reasonably possible for any constituent of the air to be discovered in addition to water vapor, nor indeed was any discovered until long after that. The chief obstacles that prevented such discovery for more than a hundred years after enough advance for that purpose had been made in laboratory technique, for that had been adequate from the beginning of the seventeenth century, were: 1) The fixed idea that all gases are alike, all just air, and that any differences between various samples are due only to greater or less modifications of one and the same thing. 2) The completely misleading and faulty concept that flame or combustion is the escape of something, phlogiston they called it, from within the burning object. 3) The failure to recognize that change of weight incident to strong heating, during combustion, or under any other circumstances, was a matter of importance or had any scientific significance whatever.

Certain studies of the air by Robert Boyle and John Mayow.—A century before any constituent of the atmosphere, except of course water vapor, was recognized and collected in an approximately pure state, and

while the faulty notions just listed were still prevalent, two people, working entirely independently, came near to finding one or more of its elements. The first of these was Robert Boyle [2] (1627-1691), a wealthy bachelor, chemist and theologian, discoverer of the fact, known as "Boyle's Law," that doubling the pressure on a gas reduces its volume by one half—of course for the same temperature. It seems very probable that Boyle would have discovered some of the constituents of the air if he had carried to completion certain experiments that he definitely listed. But there is nothing in his voluminous writings to show that he ever got them beyond the paper stage, despite the fact that, so long as his health permitted, he was a persistent worker.

The second near, even nearer, discoverer of certain of the atmospheric constituents was John Mayow [3] (1643-1679), a graduate in law at Oxford, who turned to medicine and became noted as a physician, a chemist and a physiologist. After many and well-devised experiments, Mayow concluded that the air consists of at least two portions, one that supports combustion and sustains life, and another part that does neither. The former he called "fire-air," because it keeps a flame going. He also said that it consists of "nitro-aerial particles," that is, particles in a gaseous form of the kind that make a mixture of niter (salt-peter) and charcoal, or other combustible, burn, when lighted, in the absence of air—even under water.

All this is true enough, but his proof that the air consists in part of a special constituent, different from all the rest, was not complete. He did not collect "fire-air," the gas we now call oxygen, in a practically pure form and show that it is identical with the "fire-air" of the atmosphere. However, he recognized the incompleteness of some of his arguments, and it seems likely that if he had lived a few years longer his proofs would have been perfected, and our knowledge of the composition of the atmosphere set forward almost a hundred years. But this near attainment to the goal, and also even the direct route to it, appear to have been lost sight of for nearly a century. Indeed "fire-air," that came so near to being the first constituent of the atmosphere to be discovered, except, of course, water vapor, and whose properties make it the most conspicuous, turned out to be the very last of all the major ones, save only argon.

Carbon dioxide.—The first of the permanent gases of the air of nearly constant quantity to be clearly discovered was not, as would seem most likely, either oxygen or nitrogen that together make up nearly the whole, but carbon dioxide, that is present as scarcely more than a trace—three parts in ten thousand. The investigation that led to this important discovery was begun in 1752 by Joseph Black,[4] a medical student at the

University of Edinburgh; presented as a thesis for the degree M. D. in 1754; read, with extensions, to the Medical Society of Edinburgh in 1755, and published the following year. It was not, in any sense, either a haphazard or a perfunctory piece of work. At that time alkalies were given to persons suffering with urinal calculi, and certain physicians recommended lime water for the same purpose. With the view of finding something still better, Black undertook investigations with *magnesia alba*, a form of carbonate, as we now know, or, more exactly, basic carbonate, of magnesium. On strongly heating this substance a gas is given off, and Black turned his attention particularly to that gas, or air, as all gases were then called. He tried calcining, or burning to a powder, various substances, such as limestone, that we now know to be carbonates, and studying the gas thus obtained. In the end he found that this gas is much heavier than ordinary air, that it will not support life or combustion, that it will recombine with the calx, or powder, produced by the strong heating of the original substance, and that heat will again expel it as before. This gas seems to be fixed, or somehow fastened, in the objects from which it may be obtained. Dr. Black therefore called it "fixed air." Furthermore, and this is of great importance, the calxes, like quick lime, take up identically this same "fixed air" when exposed for some time in the open, and give it off again when heated, or subjected to the action of a suitable acid. Evidently, then, as Dr. Black argued, this "fixed air" is a constituent of the ordinary atmosphere and obtainable from it.

Nitrogen.—The next advance in our knowledge of the composition of the atmosphere, the discovery of nitrogen, also was made by a medical student, Daniel Rutherford,[5] at the University of Edinburgh, and published as a thesis for his degree in 1772. At the suggestion of Dr. Joseph Black, the discoverer of "fixed air" (carbon dioxide) and then Professor of Chemistry at the University of Edinburgh, Rutherford undertook to learn what he could of the composition of the air in a closed vessel after it no longer would support combustion or life, and to determine the cause of its unwholesomeness. After burning charcoal, phosphorus, or other combustible, in a closed volume of air, as long as possible, he removed the "fixed air" (carbon dioxide), if any had been formed, by means of lime, or an alkali, all in accordance with the previous investigations of Black, and then examined the remaining gas. He showed that this residue is not ordinary air because it supports neither life nor combustion, and that it is not fixed air for the alkalies do not absorb it. He called this residue "mephitic air," because it does not support life. We now know and say that this residue had been obtained by burning out the oxygen of the confined air and then absorbing the carbon dioxide thus (in most

cases) produced. And we know, too, that this residue, this "mephitic air," was nearly pure nitrogen. But at that time oxygen had not been discovered, and of course Rutherford could not talk in terms of things and chemical reactions then unknown. He did know, however, that it was obtained by combustion in an inclosed or limited volume of ordinary air, and therefore concluded, after the philosophy of his day, that it was atmospheric air combined with, or modified by the addition of, phlogiston—a mysterious fire substance whose escape from an object commonly is manifested by flame. Nevertheless, and no matter what his ideas as to its nature, Rutherford did obtain reasonably pure nitrogen and did record some of its properties, and therefore may be regarded as the discoverer of this constituent of the air, the most abundant of all.

Oxygen.—Almost immediately after the discovery of nitrogen, the other major constituent of the atmosphere, oxygen, was independently found by Joseph Priestley, in 1774, and Carl Wilhelm Scheele, apparently before 1773. The priority of discovery belongs, it seems, to Scheele, but, on the other hand, priority of publication, by about a year, belongs to Priestley; and, as neither knew of the work of the other, each is credited with the discovery of this element, and of the fact that it is a constituent of the free air.

Priestley [6] was a preacher by occupation and a chemist for recreation. At one time, while waiting for the building of the parsonage to be finished, he had the good fortune to live next door to a brewery—good fortune, because he was induced thereby to take up the chemistry of gases, his chief avocation for many years, on which he published several volumes, and which made his name famous. In the course of this work he obtained a red powder by heating mercury in the presence of air, and then on more strongly heating this powder with a burning lens he got a gas which supported combustion much better than ordinary air. He had got oxygen, as we now know, by first burning mercury to an oxide and then decomposing that oxide by raising it to a high temperature. This element of the atmosphere he called dephlogisticated air. Combustible things burned in it readily; their phlogiston, or fire principle, rushed into it as air into a vacuum. It was air, that is, a gas, but air deprived of phlogiston. But, regardless of what he called it, Priestley had discovered a new constituent of the atmosphere, the one we now call oxygen, the one that sustains life and supports combustion.

On the other hand, Scheele [7] was a professional chemist, so completely absorbed in his subject that he had little time for anything else, whether occupation or diversion. His incentive to study the air was his wish to solve the riddle of fire, a thing essential to so many chemical processes. He found substances, such as phosphorus, which, on burning, reduced the

volume of the confined air (at the same temperature and pressure) in which they were burned by about one fifth. In all cases the remaining air would not support combustion. He also found many ways of getting a gas that would support combustion far better than ordinary air, that was heavier than the air left after combustion, and that made burnt air indistinguishable from ordinary air when mixed with it in the proportion of one volume to four, or thereabouts. The residual or burnt air, the nitrogen, essentially, as we now name it, he called "vitiated air." The other portion, the part that supports combustion, he called "fire-air," just as Mayow had called it a century before.

Scheele, like Priestley and every other chemist of his day, except the immortal and tragic Lavoisier,[8] interpreted fire phenomena in terms of phlogiston. He therefore considered "fire-air," which we call oxygen, to be a combination of phlogiston and a subtle acid substance. But whatever his opinions may have been as to its possible composition, he did find this constituent of the atmosphere, and that, too, even before it was found by Priestley.

Certain studies by Henry Cavendish.—Another student of the atmosphere who must be mentioned in connection with the discoveries of its composition was Henry Cavendish.[9] He may not generally be credited with the discovery of any one of these constituents, and yet he appears to have found nitrogen at about the same time that Rutherford did, if not earlier, by the simple process of passing the same confined air back and forth over red hot charcoal and then removing the fixed air (carbon dioxide) with an alkali. He did not publish this—he appeared always to be indifferent about publishing his investigations—but described it in a letter to Priestley in 1772, the year Rutherford published his discovery of mephitic air. A little later, 1785, he presented to the Royal Society of London an account of experiments on phlogisticated air (nitrogen) designed to test its purity. The result was that after subjecting it to a process that seemed to remove the nitrogen, and does remove it, as we now know, a small amount was left over. Beyond question he had thus obtained argon in a fairly pure state, but there is no evidence that he followed up this discovery, nor record of any one else doing so for more than a century.

Résumé to 1775.—As early, then, as 1775 the atmosphere was known to consist largely, and believed to consist almost wholly, if not quite so, of 1) water vapor, known from antiquity; 2) fixed air (carbon dioxide) reported by Black in 1755; 3) mephitic or phlogisticated air (nitrogen), published by Rutherford in 1772, and privately reported by Cavendish the same year; and 4) dephlogisticated air, or fire-air (oxygen), discovered by Priestly on Aug. 1, 1774, and by Scheele, according to his

laboratory notes, about two years earlier. Thus the problem of the composition of the atmosphere appeared to be wholly solved, and at this stage it remained stationary, except for the finding, here and there, of some variable impurity, for well over a hundred years.

Argon.—In 1894 Lord Rayleigh [10] and Sir William Ramsay [11] startled the scientific world by telling it of the presence in the air of that strange element which they called argon that in total mass exceeds by several fold the whole of the water vapor, carbon dioxide and every other atmospheric constituent all combined, except only nitrogen and oxygen. Chemists for generations had been making gas analyses by the thousands and hundreds of thousands, chemists so accurate in their work as often to detect one part in a million, and now they were told that with all their refinement of technique they had never yet detected the even more than one part in a hundred of something else, radically different, in what all along they had been pleased to call nitrogen and let it go at that. No wonder they were surprised and even skeptical.

Lord Rayleigh had been making careful measurements of the densities of different gases, and thus came upon the fact that the nitrogen of the atmosphere, or rather that residual then regarded as pure nitrogen, was a little denser than the nitrogen obtained from any one of several chemical compounds. This was capable of any one of three or four different interpretations, one of which, then regarded as the least likely, was the presence in the atmosphere of an unknown gas denser than nitrogen. At this stage Sir William Ramsay joined Lord Rayleigh in a common attack on this problem. Each tried removing from the air the then known constituents, using entirely different methods for getting rid of the nitrogen, and each found conclusive evidence of the presence in the atmosphere of an unknown gas to the extent of about one per cent. It proved to be chemically inert, and for that reason was called argon.

Other inert gases.—During the years 1895-1898, that is, immediately after the discovery of argon, four additional inert gases, helium, neon, krypton and xenon, were found by Sir William Ramsay [12] and his assistant, M. W. Travers,[13] to be constituents of the atmosphere. Helium was found as that portion of the atmosphere that is still gaseous while the rest has been chilled to liquids or solids; and the others, neon, krypton and xenon, by fractional distillations of large quantities of liquid air and liquid argon, impure, of course.

Hydrogen.—Lord Rayleigh and others also found free hydrogen always present in the air, but in amounts that varied from, roughly, one part in 1,000,000 to one in 5,000. Perhaps this gas cannot be regarded as a permanent component of the atmosphere in the same sense that oxygen and nitrogen are, but rather as an accidental and variable impurity. At

any rate it is irregularly added to the air, at least by volcanoes, and more or less irregularly removed from it.

Traces and impurities.—Varying traces of ammonia, nitric and nitrous acids and their compounds, sulphuric and sulphurous acids and their compounds, oxides of nitrogen, hydrogen dioxide, and ozone are among the innumerable things always in the air if not of it. Minute particles of sea salt from evaporating spray, fine earth dust caught up by winds, pollen of every description and spores of many kinds are among the coarser and ever present pollutions of the atmosphere from the earth beneath. Shooting stars furnish a continuous, though invisible, shower of dust the world over from outer space. Radioactive products of radium, thorium and other elements are also pouring into the atmosphere continuously and contributing to the maintenance of its electrical state. This state means, in part, around 20,000 electrified particles, or ions, per cubic inch in the lower air. The number of ions 60 miles (100 kilometers) or so above the surface is far greater, at least 1,600,000 per cubic inch as we infer from the phenomena of radio communication.

In short, while the things of the atmosphere are but few the things in it are innumerable. All these latter are relatively very small in amount, but some of them are exceedingly important. The salt particles and some others are essential to condensation and rainfall; the ammonia and other nitrogen compounds, brought down by precipitation, add much to the fertility of the soil; the ions of the high atmosphere make distant radio communication possible; and ozone, also in the upper air, shields us from that portion of the ultra violet radiation that would destroy our eyesight, as we are now constituted, and otherwise do us irreparable harm.

Percentage composition of pure dry air.—If the atmosphere is freed from all its numerous impurities and also freed from water vapor, there will remain pure dry air whose percentage composition at the surface of the earth is given in Table 1.

TABLE 1

PERCENTAGE COMPOSITION OF PURE DRY AIR

Nitrogen	78.03
Oxygen	20.99
Argon	0.9323
Carbon dioxide	0.03
Hydrogen	0.01
Neon	0.0018
Helium	0.0005
Krypton	0.0001
Xenon	0.000009

Water vapor content of the atmosphere.—The amount of water vapor in the atmosphere, or better, perhaps, the amount of the water vapor constituent of the atmosphere, varies from scarcely more than a trace, at extremely low temperatures, to at least 5 per cent by volume on the hottest and most humid days. At and above the height of 5 miles, say, the amount of water vapor always is small, even when saturation obtains, owing to the very low temperatures at these levels. That is, water vapor is confined almost wholly to the lower atmosphere. Its average mass, the world over, is such that if it were all condensed it would be the equivalent of a layer of water about one inch deep over the entire earth.

Forms of oxygen and their distribution.—Oxygen also is peculiar in its distribution, and it occurs in three different forms. All but about one part in 400,000 is ordinary oxygen, or, in the language of the chemist, diatomic oxygen. Triatomic oxygen, or ozone, occurs almost exclusively beyond the highest clouds, its greatest density being, apparently, at the levels of 20 to 30 miles (30 to 50 kilometers). It occurs in the lower air only as a mere trace, if at all. In the highest air, 60 miles (100 kilometers) and more above the surface, the oxygen appears to be monatomic, according to the spectrum of the aurora.

Importance of ozone.—Every one knows that life would be impossible if there were no ordinary or diatomic oxygen in the atmosphere, and that without it nearly if not quite all vegetable life also would be impossible. But it is not so well known, though equally true, that animals, including the human species, could not exist as now constituted if the air did not contain a small amount of triatomic oxygen or ozone. And yet, paradoxical as it may seem, if ozone were just a little more effective in its goodness, again life, as now constituted, could not last. These surprising facts come about in this way: The radiation from the sun includes not only every color, that is, the whole of the visible spectrum, but also extends indefinitely beyond the red into the long-wave length invisible region, and likewise well beyond the limit of the violet. But in this ultra violet portion of the spectrum the radiation from the sun ceases far short of the limit to which that from an electric arc, for instance, can be followed; and ceases, not because no radiation beyond that limit is given out by the sun, but because it is absorbed by the ozone in the upper atmosphere. Now, much of that particular radiation which ozone absorbs is destructive to the eye, and when intense probably injurious to other tissues as well. On the other hand, it also absorbs a large part of the radiation that is effective in preventing rickets, but, and this is of the utmost importance, it does not absorb quite all of this antirachitic portion. Enough is let through to keep us in proper health. This particular limiting, then, of the solar radiation that reaches the earth is amazingly located. If it was a little farther

out in the ultra violet, eyes, as now constituted, could not have developed; and if a little nearer the visible, again, animals, as they now are, could not have come into being. Of course ozone, in moderate amount, presumably was in the air long before there were animals of any kind to be affected by its action on solar radiation. This radiation, therefore, was not adapted to them, but they developed in adaptation to it; nevertheless, the fit is so close and of such a surprising nature as to give us decided pause for thought.

Origin of ozone.—Ozone is produced by the action of extreme ultra violet light on oxygen, and therefore at great heights; a little where the oxygen is very rare, then more and more with decrease of height and increase of the oxygen supply until by absorption the ozonizing rays are considerably enfeebled. As the density of the oxygen increases, the intensity of the effective radiation decreases, hence the rate of production of ozone by this process must be very slow in the outermost air, increase to a maximum with decrease of height, and then rapidly fall off at still lower and lower levels. The total range of ozone, top to bottom, probably is not less than 100 miles (160 kilometers), with its maximum concentration around 25 miles (40 kilometers), perhaps, above the surface,[14] and yet all told, and despite its great importance, it is the equivalent of a layer of this gas only about one tenth of an inch (2.5 millimeters) thick at atmospheric pressure and room temperature.[15] Ozone is produced also by electric discharges through oxygen. Lightning certainly produces some ozone, and it may be that the auroras form it too, though the great heights at which they occur, 60 miles (100 kilometers) and more above the earth, seem to render this conclusion doubtful.

Distribution of ozone.—As already stated we know that at most there is only a trace of ozone in the atmosphere up to the level of the highest clouds, and that it exists to an appreciable extent at considerably greater heights. But this is not all. The total amount of ozone vertically over one to the limit of the air appears to increase, in general, with increase of latitude; to be greater during winter and spring than during summer and fall; and to be greater in winds from high latitudes than in those of the opposite direction.[15] This complicates the problem of the origin of ozone. If it is produced by ultra violet radiation alone, why should it not be most abundant in tropical regions, and elsewhere in the summer time? If we surmise that it is produced largely by auroral discharges, how, we ask, can these discharges, 60 miles (100 kilometers) up and more, reach a sufficient supply of oxygen? And if there is enough oxygen at these great heights how can the ozone subsequently become concentrated at the intermediate levels? These are some of the questions to be answered by future observations and studies.

Upper and lower atmosphere.—The composition of the lower atmosphere up to at least 6 or 7 miles (10 or 11 kilometers) in temperate regions, and 8 or 9 miles (13 or 14 kilometers) within the tropics, is well known, except in respect to condensation nuclei and certain impurities. Throughout this region, too, the percentages of the several gases, except water vapor, is practically constant, owing to their continual mixing incident to convection and turbulence. Somewhere in the upper air, however, the percentages of the lighter gases must increase and those of the heavier decrease under the action of gravity, since at these levels vertical convection is very feeble. But as the upper air grows thinner, the more and more rapidly does our knowledge of it become less. We know that its outermost portion is the very extensive region of the aurora, hundreds of miles, or kilometers, thick; that near the base of the auroral region there is enough ionization to make wireless communication around the world entirely practicable; and that far below this Heavyside layer, in turn, and yet well within the upper air occurs most of the ozone, the triatomic oxygen that indirectly is so vital to all terrestrial life. We seldom give any thought to this upper air, but it is so important that we really must know more about it.

Mass of the atmosphere and of its constituents.—From the known percentages of the several constituents of dry air, given above, their molecular weights, and various other pertinent facts, such as the amount of water vapor present, height of the barometer, volume of land above sea level, and distribution of temperature with height, it is easy to compute the approximate mass of the atmosphere as a whole and of each of its several gases. The results are given in Table 2, in which the factor 10^8 means: Add eight ciphers, or multiply by 100,000,000.

TABLE 2

MASS OF THE ATMOSPHERE AND OF ITS CONSTITUENTS IN TONS (2,000 POUNDS)

Substance	Volume per cent dry air, at surface	Total mass	
Total atmosphere	.....	56,328,000 x 10^8	tons
Dry air	100.00	56,181,850 " "	"
Nitrogen	78.03	42,684,725 " "	"
Oxygen	20.99	12,782,647 " "	"
Argon	0.9323	682,125 " "	"
Water vapor	.....	146,150 " "	"
Carbon dioxide	0.03	23,874 " "	"
Hydrogen	0.01	1,423 " "	"
Neon	0.0018	759 " "	"
Krypton	0.0001	141 " "	"
Helium	0.0005	88 " "	"
Ozone	0.00006	33 " "	"
Xenon	0.000009	19 " "	"

Since the values in this table were determined more or less independently it could not be expected that the percentages found of the constituents would add up exactly 100, nor that the sum of the computed masses of the several parts would precisely equal the mass of the whole. These deviations, however, are very small—probably within the present limits of experimental errors.

The numbers here given that express the masses of the atmosphere and its several constituents are useful as quantitative values and for exact comparisons, but so great, even though in terms of tons, that we can form no distinct conceptions of them—they are just awfully big! A clearer idea may be obtained from the fact that the total mass of the atmosphere is the equivalent, roughly, of that of a block of granite a thousand miles (1,610 kilometers) long, a thousand miles (1,610 kilometers) broad and a half mile (0.8 kilometer) thick; while the least abundant of the constituents, xenon, if loaded on cars, 19 tons (17,240 kilograms) to the car, would freight a train which would reach forty times around the earth along a great circle, and which, traveling twenty miles an hour (9 meters per second), would be six years in passing any fixed point on the road.

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CHAPTER II

METEOROLOGICAL DATA AND METEOROLOGICAL CHANGES

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INTRODUCTION

This chapter is concerned with what are commonly known as climatic changes of a more or less enduring character; such changes may be read from the geologic record and to a less extent from other sources, as will be indicated later. The discussion naturally falls into not more than four periods as follows: 1) the geologic record up to, say, near the dawn of historic time; 2) a more or less shadowy period from the end of the geologic past to the beginning of historic time; 3) from the end of 2) to the beginning of the instrumental period, say the end of the eighteenth century; 4) the present, from the beginning of the nineteenth century. These periods will now be considered in order.

NON-INSTRUMENTAL PERIOD

(1) *The geologic period.*—The geologic evidence shows that from the beginning of geologic history the earth periodically, or, perhaps it is better to say at irregular intervals, has undergone widespread glaciation with its attendant lowering of the temperature and destruction of all forms of life in those regions most intensely glaciated. Relatively warm periods have been interspersed with the different epochs of glaciation. The duration of the warm periods as well as that of the extremely cold periods is uncertain. The evidence relied upon by geologists consists, in the main, of the distribution throughout the world of glacial till, the tillites of the geologist, boulders, and the remains of fossil plants and animals. Geologic evidence alone presents some difficulties in elucidating climatic changes, since it is often impossible to determine with much precision the date when the deposits were laid down.

The study of the fine seasonally banded clay deposits (varves) laid down in quiet lakes and fiords in the glaciated regions of Europe and North America by De Geer and Antevs has led to greater precision in dating the retreat of the glaciers at different stages of their history. From such a study the Swedish geologists have placed the end of the glacial period in Sweden at about 6500 B. C.[1]

(2) *From the glacial to the historic period.*—From 6500 B. C. to the beginning of the Christian era the evidence of climatic change is more

or less uncertain, although the author just cited has summarized that which is available for Europe and Asia.

Ellsworth Huntington has utilized field studies in Asia and North America for evidence of progressive, or, perhaps a mode of, desiccation that occurs in the form of pulsations which endure for a century or more and are then followed by a swing in the opposite direction. He says: [2]

There is strong reason to believe that during the last two thousand years there has been a widespread and pronounced tendency toward aridity. In drier regions the extent of land available for pasturage and cultivation has been seriously curtailed; and the habitability of the country has decreased. . . . . Moreover, in both the drier and the moister regions the change of climate does not appear to have been all in one direction. After a period of rapidly decreasing rainfall and rising temperature during the earlier centuries of the Christian era, there is evidence of a slight reversal and a tendency toward more abundant rainfall and lower temperature during the Middle Ages.

(3) *The historic period.*—There is yet another form of evidence, found in the instrumental period, that seems to point to progressive climatic change, suspected but not as yet definitely proved; viz., progressive change so slow as scarcely to be noticeable in instrumental records, such, for example, as has been suggested is taking place in southwest Africa, where it has been stated that precipitation is growing less and less.

(a) *Pulses in rainfall.*—Since writing the foregoing it has come to our attention (Nature, February 2, 1929) that Professor E. H. L. Schwarz, who in 1918 published his well-known scheme of diverting water from the Zambesi into the great depressions of Lake Ngami and the Kalahari, has but recently completed a reconnaissance of his proposals and published the results in a work entitled The Kalahari and its Races: being the account of a journey through Ngamiland and the Kalahari (London, 1928), in which he remarks (page 13), "A country that had resigned itself to the condition of permanent drought was for a time gladdened by the sound of rippling water on all sides." Prof. Schwarz found the lake again occupied by water and submits a summary of its variations.

In 1760 it was dry, then followed a period when it was a great lake, from 1813 to Livingston's visit in 1849, when it had begun to decrease; from 1854 until 1861, when it held some shallow water surrounded by reeds; and from 1896 until 1922, when there was no water and the lake was a dry plain.

The restoration of Lake Ngami is regarded as evidence of a cyclic climatic change that has run its course in a little more than 100 years.

(b) *The evidence of tree-rings.*—Variations in the width of the annual growth rings of certain trees, for studies of which we are indebted to Dr. A. E. Douglass, Director Steward Observatory, University of Arizona, Tucson, Ariz., show a correlation with rainfall of the few years available during historic times and this relation has been used by several

writers as an index of the climate of the past. Douglass, in his original studies, used the rings of the western yellow pine. He says of this tree:[3]

The western yellow pine is perhaps the best tree for climatic studies, on account of its precision and length of record and its wide distribution. It is normally a dry-climate tree and does well in a sandy soil, for its thick bark prevents evaporation from the trunk and thus enables it to live when other trees could not survive. Thus it endures relatively trying conditions and has little competing vegetation, so that the Arizona forest is said to be the largest "pure" stand in this country. . . . . Its age is very favorable, reaching over 500 years. It is commonly free from burns and defects and its rings are very readable. The immense area over which the yellow pine grows adds to its value in this study, as its use avoids the complexities arising from the use of different species. For all these reasons it is considered the standard tree.

In order to carry the record back beyond the age of the yellow pine Douglass extended his studies to include the Sequoia (*Sequoia gigantea*) for, because of its great age, it has a fundamental feature of great importance, viz., cross-identification over large areas, and in this character the influence of climate is recognized. The ring-growth of the big Sequoia is not so sensitive as in yellow pine, and perhaps any individual tree is a little less certain to identify with its neighbors; but cross-identification is very sure in that species, and extends through all the mountain sequoia groves from Calaveras on the north to Springville on the south, a distance of 200 miles.

The great age of the tree gives it a second fundamental value. It is astonishing for example to find over considerable areas identifiable rings near 1000 B. C.

Lack of space forbids a detailed description of the corrections applied to measures of tree-rings and the care exercised to exclude personal bias in reaching the results.

In discussing the previsional value of tree-ring data Dr. Douglass holds closely to a conservative course and indicates quite clearly in five paragraphs certain cautions that should be observed in interpreting the data. We have condensed these in the short paragraphs which follow.

First caution.—Arizona trees respond closely to a definite weather element, the rainfall, but in moist areas this direct response decreases and even disappears; hence the caution that one must not assume relationships similar to those of Arizona in any given place until that place has been thoroughly investigated.

Second caution.—This caution emphasizes the fact that cyclic changes are not understood and that until the physical causes are known it is impossible to say how long a mechanical repetition will last, for it may break down at any time.

Third caution.—This caution concerns the effect of splitting of cycles and the complexities that arise therefrom.

Fourth caution.—This caution has to do with the results of the interference between cycles of short length and those of longer period.

Fifth caution.—It is pointed out that each of the western zones investigated has a central homogeneous area whence variations spread outward in all directions, so that in some cases a region may come under the influence of one or more central homogeneous areas, in which case, prediction will be difficult.

The extension of the Flagstaff area synthetic curve into the future is ventured by Dr. Douglass in the following words:

Flagstaff area synthetic curve.—The mean curve covering the area from the Grand Canyon to the Rim shows very excellent similarity to the individual curves composing it, but many of the short periods have disappeared and multiples of 7.0 years are left prominent, 21 years being by far the strongest. Residuals between the synthetic curve and the real growth-curve show a 9.4 cycle in the latter part of the eighteenth century. Crests are too high (in the natural curves) at 1793 and 1891 and the minima at 1847 and perhaps 1880 are too low to be accounted for by the synthetic curve. The 7-year cycle was almost absent from 1880 to 1905. Yet on the whole there is a good deal of similarity. The prolongation of the synthetic curve shows a small depression near 1927 and deeper ones at 1942 and 1947. The interval between the 1930's has high ordinates with an unimportant depression at 1933. It is possible that the 1947 depression may resemble the one of 1847 and be rather extreme. During the 1950's the curve is again high. High crests occur at 1937 and 1953. It is not expected that this is entirely right, but details are given here in order to assist ultimately in finding the true variations.

The tree-ring data appear to have certain limitations as indicated by the author in the foregoing excerpts. With these in mind and the further one that the curves are plotted from smoothed data, the effect of the smoothing being to reduce the amplitude of the oscillations and slightly to displace the epochs of maximum and minimum, the student may accept them as being indicative of climatic changes in the arid southwest of the United States. Whether they are of world-wide application the writers are not prepared to say; caution No. 1 would, however, appear to negative such a conclusion.

The Brückner cycle.—Passing now to a consideration of a large mass of miscellaneous non-instrumental data all of which were collected and discussed by the late E. Brückner about 40 years ago, we may preface our remarks by the observation that the so-called Brückner cycle of climatic oscillations might more properly be designated as the Brückner years of heavy and light rainfall, etc., since the regularity of occurrence does not warrant the use of the term "cycle." We present below a rather detailed statement with reference to what is commonly known as the "Brückner" cycle, which is extracted from a paper that was presented by the writer before the joint meeting of the Association of American Geographers and the American Meteorological Society at Philadelphia, December 30, 1926.

In 1890 Dr. E. Brückner, professor of geography in the University of Berne, announced the discovery of a climatic cycle of an average length of thirty-five

years. The quantity of evidence submitted was overwhelming and the reality of the cycle has not been seriously questioned. Some modern writers, however, have attributed to the cycle much more stability than it really possesses, or than was ever claimed for it by its discoverer.

Brückner's work[4] was published in Penck's Geographische Abhandlungen, Band IV-Heft 1, in 1889, and in a separate volume in the succeeding year.

One who has examined the original presentation must be impressed by the very great mass of material assembled and discussed. But whether it is humanly possible for any single individual, however well equipped he may be, to critically examine, interpret, and appraise the meteorological records from all parts of the earth is another matter.

The great bulk of Brückner's material is drawn from European sources and is presented in the following order:

(1) Secular variation of the Caspian Sea.
(2) do of lakes and seas without outlet.
(3) do of river heights.
(4) do of precipitation.
(5) do of atmospheric pressure.
(6) do of temperature.
(7) do of times of grape harvest in Europe, the occurrence of severe winters, the advance and retreat of glaciers, etc., all or the greater part of which is found in European records.

(a) *Average length.*—This is said to be thirty-five years, although Clough[5] assigns a mean value of thirty-six years. Each cycle is composed of a cool-humid half and a warm-dry half; the yearly temperature and precipitation, for example, increases to a maximum and then decreases to a minimum, to be sure in an irregular manner, but in such way as to be recognized in five-year means. The time between two maxima or two minima may vary between twenty and fifty years.

(b) *Amplitude.*—The amplitude of the temperature oscillations counting from the highest to the lowest point reached in the five-year means is small, not exceeding one degree Centigrade, and in the term of years 1836-71 it was only half a degree C. The variation in precipitation for five-year groups of means for the years 1731-1881 was from 9 per cent above the normal to 8 per cent below. Much greater local variations both of temperature and precipitation occur in various parts of the world; but in the grouping of a large number of stations in a single mean the tendency is to reduce the extreme values which are found for individual stations.

Brückner's results, in so far as they are derived from systematic meteorological observations, are based essentially upon the records available for the fifty-six years between 1830 and 1885 inclusive.

The number of useful temperature and precipitation records prior to those years was inconsequential.

Brückner found in his studies of climatic oscillations certain large regions which he has called "regions of regular oscillations"; in other regions he found a distribution the reverse of that just mentioned and these were classed as "regions of permanent exception"; and in still other regions he found a distribution that for a time followed his "regular" distribution and then suddenly reverted to something different. These he classed as "regions of temporary exception." In his later writings, however, he claims that the oscillations are world-wide.[6]

(c) *Argument against the Brückner cycle.*—The strongest argument that can be adduced against the usefulness of the Brückner oscillations is the fact that precipitation and temperature fluctuate in somewhat the same manner as postulated by Brückner but in much shorter periods. For example, in the United States during the nineteenth century, there were as many as nine distinctly warm and distinctly cool periods or one to every ten years. The amplitude of these short-period oscillations was equally as great as and sometimes greater than that which took place at the time of maxima and minima of the Brückner 35-year oscillations. In the case of precipitation it is well known that there are more dry than wet years, and further, that a wet year may be and frequently is intercalated in a series of dry years. For this and other reasons it is held that the so-called Brückner cycle has no previsional value.

THE INSTRUMENTAL PERIOD

Meteorological instruments.—An excellent description of meteorological instruments employed by the U. S. Weather Bureau has been prepared by Covert.[7] These instruments represent fairly well the equipment to be found in meteorological observatories the world over, except that in many small observatories only a few of those here enumerated will be found.

The instruments may be divided into two classes, namely, recording and non-recording. Every first-class station has a full set of non-recording instruments of standard type, since they may be read more accurately than automatic instruments can record. They include a mercurial barometer; an alcohol thermometer for indicating the minimum temperature of the day; a mercurial thermometer for indicating the maximum temperature of the day; a dry and a wet-bulb thermometer, usually mounted so that they may be whirled about a horizontal axis to obtain from the first named the temperature of the air and, by computation, from this temperature and the depression of the wet-bulb temperature, the dew-point temperature. This latter makes possible the determination of the absolute and the relative humidity of the atmosphere. All thermometers are usually exposed in an instrument shelter to protect them from the effects of radiation.

A funnel-shaped rain gage in which to collect and measure the rainfall, and a wind vane to indicate the direction of the wind, are two instruments that seem to have been first used a few centuries before the Christian era.

Early meteorological records.—Although the barometer and thermometer were invented about the middle of the seventeenth century, their use in systematic measurements of pressure and temperature did not begin or become general until about the middle of the nineteenth century and

even then the network of meteorological stations in the majority of countries was very thin.

At the beginning of the nineteenth century worth-while series of meteorological observations were being made at but twelve points in Europe, the earliest in point of beginning being the series of rainfall measurements that was started at Padua, Italy, in the year 1725. By far the greater number of European stations began observations in the second half of the nineteenth century.

In North America observations had begun at 10 stations by the end of the eighteenth century, but only half of that number had begun soon enough to accumulate a set of observations of any value, and three of these were bunched in New England; the remaining two were located, one in Philadelphia, Pennsylvania, and the other at Charleston, South Carolina. At the beginning of the nineteenth century there were all told but 12 stations in Europe and 5 in North America that have yielded dependable results. The number increased considerably up to the middle of that century but the great increase in all countries took place in the second half of the nineteenth century. The distribution of the observing stations fell far short of that which could be wished, and there are still vast areas in Asia and Africa that are not represented on a world map.

The foregoing paragraph means that the modern instrumental period is essentially comprised within the last one hundred years and that for a large part of the earth's surface less than 100 years of dependable observations are available; the record of sunspots, we may say, extends back to 1749 and it has been extrapolated still farther back. A notable piece of work in extending backwards the record of cold winters in northwest Europe is that of Dr. C. Easton, whose historical studies have been published under the title Les hivers dans L'Europe occidentale. This publication carries the record back to 396 B. C.

Modern meteorological records.—A regular meteorological observation is made at a stated time at all observing stations included in a wide area. In the United States and Canada the time is 8 a. m. and 8 p. m., 75th meridian time. It consists of eye readings of the barometer, the rain gage and the thermometers enumerated above, and of the wind direction and velocity from recording instruments. A careful record is made of the kind, amount and direction of movement of all cloud forms, and of the state of the weather.

Cloud records.—A great deal of attention is given the cloud record. The movement of the clouds at different levels is an indication of the air movement at those levels, and the presence of the cloud indicates that the air is saturated with moisture at the place where the cloud forms. The kind of cloud indicates something of the processes taking place in the

3

atmosphere to bring about saturation. Thus a cumulus cloud indicates vertical convection, the cloud forming at the top of a vertical convection current. A sheet of stratus clouds may result from forced convection, where a warm current rises to overrun a cold one, or from the mixing of air at the boundary of a warm moist and a cold current, the latter perhaps already nearly saturated. Halos form in cirro-stratus clouds, which occur at such great altitudes that the moisture they contain is in the form

Fig. 1.—Tufted forms of cirrus clouds. Photographed by F. Ellerman.

of snow crystals; coronas and rainbows are formed by water droplets in comparatively low clouds.

Cloud Classification.—This has been accomplished through international cooperation and agreement and an international commission is at the present time engaged upon a revision of the classification now in general use, which was first issued in 1896 and revised in 1910.

This classification recognizes ten principal types, based upon the form of the clouds, as follows:

1. Cirrus (ci.). See Figure 1.

Detached clouds of delicate and fibrous appearance, often showing feather-like structure, generally of a whitish color, and sometimes arranged in parallel belts, which cross the sky in great circles, and by an effect of perspective appear to converge towards a point on the horizon.

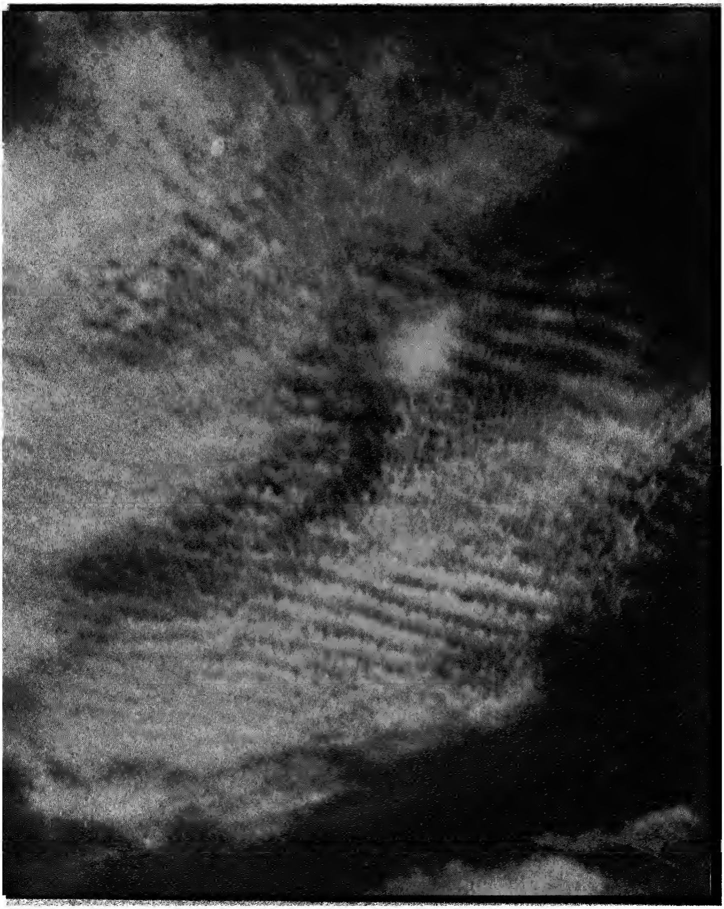

FIG. 2.—Cirro-cumulus clouds, overhead. Photographed by A. J. Henry.

2. Cirro-stratus (Ci.St.)

A thin whitish sheet of clouds, sometimes covering the whole sky and giving it a milky appearance, and at other times having the appearance of a tangled web.

3. Cirro-cumulus (Ci.Cu). See Figure 2.

Mackerel sky. Small globular masses, arranged in groups and often in line, and showing little if any shadow.

Classes 1-3 form at high levels (5-12 km.).

4. Alto-Stratus (A.St.).

A thick sheet of gray or bluish color, sometimes resembling thick Ci.St.

5. Alto Cumulus (A.Cu.). See Figure 3.

Fig. 3.—Alto-cumulus clouds. Photographed by Wilson-Barker.

Largish globular masses, white or gray, partly shaded, arranged in groups or lines, and often closely packed.

Classes (4) to (5) are intermediate in height (3 to 6 km.), classes (6) to (10) are all lower clouds (below 3 km.).

6. Strato-cumulus (St.Cu.).

Large globular masses or rolls of dark clouds often covering the whole sky, especially in winter.

7. Nimbus (Nb.).

A thick layer of dark clouds, without shape and with ragged edges, from which steady rain or snow falls.

NOTE.—Many meteorologists object to making this a separate class. They would prefer the suffix *N* after the abbreviation for the cloud class. Thus St.N. would signify a stratus formation from which precipitation was occurring.

8. Cumulo-nimbus (Cu.Nb.). See Figure 4.

The thunder cloud; shower cloud. Heavy masses of cloud rising in the form of mountains, turrets or anvils, and having at its base

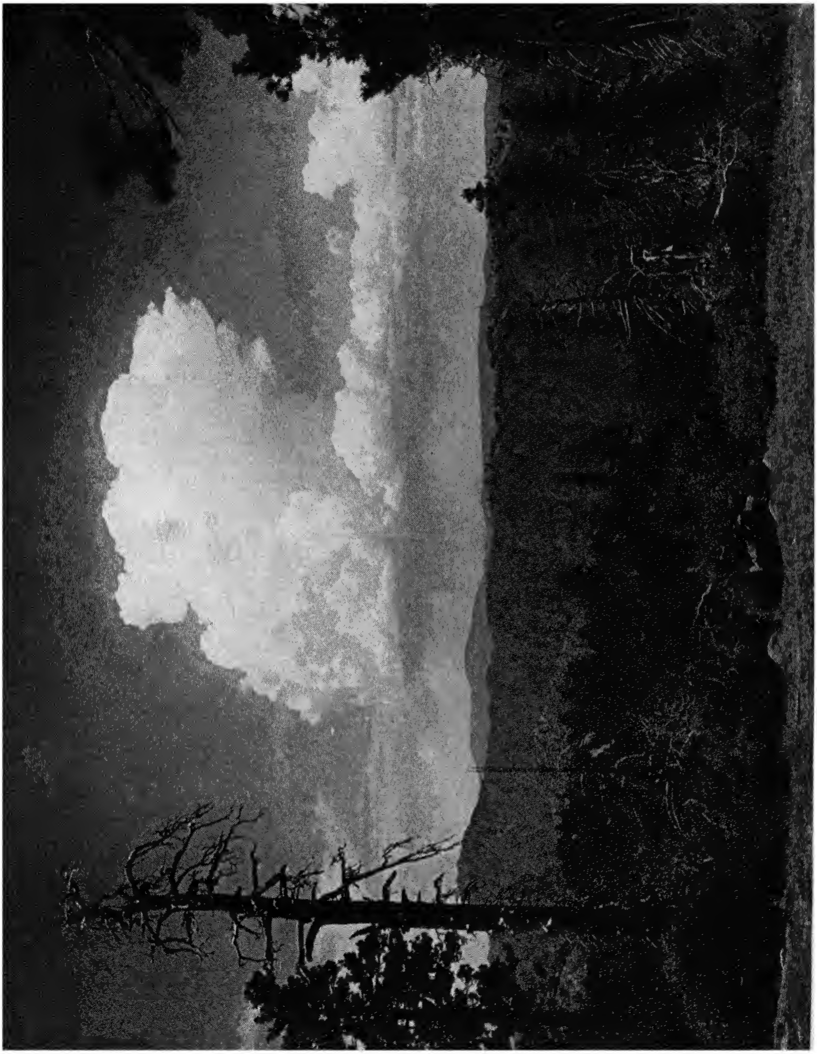

FIG. 4.—Cumulo-nimbus cloud. Photographed by F. Ellerman.

a mass of cloud similar to nimbus; generally surmounted by a sheet of "false cirrus."

9. Cumulus (Cu.). Wool-pack clouds. See Figure 5.

Thick clouds of which the upper surface is dome-shaped and exhibits protuberances while the base is horizontal.

10. Stratus (St.). See Figure 6.

Fig. 5.—Cumulus clouds. Photographed by W. M. Lyon.

Fig. 6.—Stratus clouds at two levels; one practically on the ground. Photographed by A. J. Weed.

A uniform layer of cloud resembling a fog, but not resting on the ground.

For the exact determination of height and movement of clouds, measurements with theodolites and nephoscopes are necessary. Photographic cameras mounted on range finders are also used.

Daily record.—A *daily record,* or *Journal,* containing an account of special phenomena, such as halos, auroral displays, frost, fog, thunderstorms, hail, etc., completes the record of the weather.

Summaries and means.—From these records, daily, weekly, and monthly means and summaries are formed, but the weekly means are coming more and more into use for agriculture and general climatic studies.

Investigation and mapping of world weather.—Several notable contributions have been made to this subject and these have been preceded by the mapping of the weather for the whole or some part of the globe. The following list of world charts has been excerpted from Sir Napier Shaw's Manual on Meteorology, 1: 287.

(a) *Daily charts of the oceans.*

1861. Synoptic charts of the Indian Ocean for the months of Jan., Feb., and March, Mauritius.

Sept., 1873–Nov., 1876. Hoffmeyer's charts of the North Atlantic.

1877-87. Weather charts of the Northern Hemisphere, published by U. S. Signal Service.

1880-1910. Daily synoptic charts of the North Atlantic, published by the Danish Institute and the Deutsche Seewarte, jointly.

Aug., 1882–Aug., 1883. Daily charts of the North Atlantic, published by the Meteorological Office, London.

1900-14. Dekaden Bericht of the Deutsche Seewarte—daily charts of the North Atlantic and adjacent continents.

1912-14. Daily charts of the North Atlantic, issued with the Weekly Weather report of the Meteorological Office.

1914. Daily weather charts of the Northern Hemisphere, U. S. Weather Bureau.

1923. Daily charts of the Pacific Ocean, published by the Imperial Marine Observatory, Kobe, Japan.

Daily charts of the South Atlantic Ocean for the period of twelve consecutive days, 1909, February 20 to March 3, are given on the back of the Monatskarte für den Nordatlantischen Ocean for February, 1911.

The third entry in the above list shows that a little more than fifty years ago the Army Signal Service of the United States began the preparation and issue of daily weather charts for the Northern Hemisphere and continued that issue at very great expense for ten years. The results of this pioneer effort were printed *in extenso* and the ponderous volumes containing the observations and charts still encumber the shelves of most of the great meteorological libraries. There they serve not so much as

valuable working material as a reminder of a worthy effort that was started about one hundred years before its appointed time. The financing of such an undertaking was possible under the conditions which obtained fifty-odd years ago, but there is little probability that a similar program could be carried on at present by any single meteorological service.

The institution of the Réseau Mondial under the auspices of the International Meteorological Committee and printed as Part V of the British Yearbook was a forward-looking step in the interests of world meteorology. The latest issue of this publication is for the year 1921, and although the authorities having the publication in hand are alive to the need of more prompt printing, it is quite evident that unless the contributing meteorological services of the world speed up the sending of their meteorological summaries to the London Meteorological Office, there is little hope of earlier printing. Thus far 12 issues of Réseau Mondial have appeared (1910-21).

Observational material available.—A fairly large body of observational material was not available for study until near the close of the nineteenth century; since that time it has been accumulating rather steadily, but not in proportion to the rate between 1850 and 1900. There were 437 stations utilized in the preparation of the Réseau Mondial for the year 1920. These stations were distributed as follows:

Number of stations having normals:

	Pressure	Temperature	Precipitation
Northern Hemisphere	242	257	259
Southern Hemisphere	97	98	108
Total	339	355	367

About 8 per cent of the total number of stations do not yet have normals. The Réseau Mondial gives, of course, only current meteorological data. Recognizing the need for a convenient summary of the monthly mean values from the longest established stations of the world, means to assemble and publish them in a single volume was provided in 1923 as recited in the next paragraph.

(a) *World weather records.*—The International Meteorological Committee, at its meeting in Utrecht in September, 1923, adopted a resolution to the effect that the publication of long and homogeneous series of observations in the form of monthly means of pressure, temperature and rainfall would be of the highest importance for the study of the general circulation of the atmosphere. The execution of this resolution was delegated to:

Sir Gilbert Walker, for stations in Asia;

Professor Exner, for stations in Europe;

Mr. Clayton, for stations in the Americas; and

Dr. Simpson for stations in Africa, Australia and the oceans.

The work was put through the press by Mr. H. H. Clayton and the expense of publication was borne by Mr. John A. Roebling, a citizen of the United States.

The volume contains the record of monthly means of pressure and temperature, and the total monthly precipitation for a total of 402 stations; but the records of the individual stations are not uniformly complete for each of the three elements named. Some of them lack pressure, some temperature, and a few lack both; moreover, for practically all of the network of stations belonging to the old Russian Empire the records start with the year 1881, whereas many stations were in operation at least ten years earlier, and the station formerly known as Saint Petersburg should have a record extending back into the eighteenth century. The longest rainfall record of modern time, that of Padua, Italy, for reasons not known, is omitted. These statements are not made in a critical spirit, but by way of reminder of what must be obvious, viz., that a vast amount of preliminary work in reducing and making homogeneous long records of monthly means, is necessary. If some one would do for Eurasian pressure records what Bigelow did for those of North America, full use might be made of the volume of world records under discussion.

The Challenger expedition, 1872-1876.—Although a number of short exploratory sea voyages were available about the middle of the nineteenth century it was not until the last half of that century that a serious and sustained effort was made to investigate the meteorology of the oceans. Owing to representations of the council of the Royal Society of Great Britain, the government of that country detailed H.M.S. "Challenger" a wooden corvette of 2,306 tons for the purpose of prolonged oceanic exploration not only of the atmosphere above the sea but of the sea itself, its currents, temperature, salinity, etc. The vessel left Portsmouth in December, 1872, and during the ensuing four years was continuously engaged in a voyage of scientific discovery. Meteorological observations were made hourly or two-hourly throughout the voyage. The data thus accumulated were turned over to Mr. Alexander Buchan who devoted about seven years of his life to reducing and discussing them. At that time (early eighties) there were available as bearing on world meteorology the following-named contributions:

Dove: On the distribution of temperature over the Globe, 1852, and for the Northern Hemisphere, 1864.

Buchan: On the mean pressure of the atmosphere and prevailing winds of the Globe. Trans. Roy. Soc. Edin., vol. xxv.

Coffin and Woeikof: On the winds of the Globe. Smithsonian Contributions to Knowledge, 1875.

The results of the Challenger Expedition were published in 1889 and for the first time it was possible to show the world-wide distribution of the three elements, temperature, pressure and prevailing winds on a single map, all for the 15 years ending with 1884. Two sets of charts, one for the globe the other for the northern hemisphere on a polar projection, are given in the report and we may say that the publication of these investigations laid the corner stone for the investigations that followed.

World meteorology.—Hildebrandsson and Teisserenc de Bort (1907) in their work Les bases de la meteorologie dynamique, made an important contribution to world meteorology and Hildebrandsson in a later work, Quelques recherches sur les centres d'action de l'atmosphere (1914), plays the part of a pioneer in establishing the pressure relations that exist between distant regions, using for that purpose the method of parallel curves. Later, these relations were investigated by the statistical method of correlation, chiefly by Sir Gilbert Walker while he was Director of Indian Observatories. Walker worked out the correlation coefficients between the weather elements at a large number of places in various parts of the globe and applied some of the results in successfully predicting the monsoon rains of India, as elsewhere indicated.

There has been no lack of effort on the part of investigators to discover the fundamental principles of the atmospheric circulation and to relate them to the orderly sequences of the weather as we shall indicate on a subsequent page.

Although the idea of international cooperation has been a fixed and guiding principle since the meeting of the first international congress at Leipzig in 1872, the conditions that make for really effective cooperation have been wanting by reason of differences of environment and of local needs which interpose an obstacle to effective cooperation; and, moreover, the members of that and subsequent congresses not having official standing with their home governments have been without power to guarantee the necessary financial backing to carry into effect plans looking to a more complete cooperation than now exists.

The meteorology of the free air.—Up to this point in the discussion, no mention has been made of meteorological observations in the free air above the immediate ground surface.

So far back as in 1784 the minds of scientific men had turned to the manned balloon as a means of securing information regarding the conditions of the air strata some distance above the ground. In that year Dr. James Jeffries, an American medical student temporarily sojourning in London, paid one hundred guineas for a seat in a balloon piloted by a Frenchman, M. Blanchard.[8] In that ascent Jeffries collected air samples at various heights and made meteorological observations.

In sketching an outline of the early history of ballooning in the interest of meteorology, one naturally thinks of the voyages made by the pioneer of ballooning, John Wise, who in June, 1843, was caught in a hailstorm over Carlisle, Pennsylvania. His experience in that storm served as definite proof of the method of hail formation and removed much of the vagueness of current theories then prevalent. One also thinks of the pioneer work of Glaisher and Coxwell in England during the years 1862-66. At the instance of the British Association for the Advancement of Science twenty-eight ascents were made by these men. After this series of ascents the practice of scientific ballooning lagged for a time but was revived by the Germans, with better equipment and a better understanding of the objects to be accomplished, in the eighties and nineties of the nineteenth century.

The German program consisted of the making of many balloon ascents under varying conditions, the results of which have been published under the title Scientific Air Voyages Carried out by the German Union for the Promotion of Aeronautics in Berlin, with the cooperation of O. Baschin, W. von Bezold, R. Bornstein, H. Gross, V. Kremser, H. Stade, and R. Süring, put through the press by Richard Assmann and Arthur Berson in three quarto volumes, Vieweg und Sohn, Braunschweig, 1899-1900.[9]

The hazard of life incident to the use of balloons as a means of free air research led to the practice of sending up small unmanned paper balloons by Hermite and Besançon in 1892, and thus the ballonsonde (sounding balloon) came into existence. This method was further developed and successfully used by M. L. Teisserenc de Bort, in his observatory at Trappes, France. The sounding balloon carries automatic recording instruments and a parachute to bring it safely to earth again. It has been invaluable in securing data of the free air up to 20 and 30 km. (12 to 19 miles).

Kites also have been employed in investigating the free air but their use is conditioned upon the state of the weather, especially upon the speed of the surface wind. Although they have served a useful purpose in exploring levels up to say 4 or 5 km., by far the greater knowledge of the free air has been gained by the use of sounding balloons.

The great advances in civil aviation since the close of the World War have necessitated the collection and distribution of a large amount of local as compared with general meteorological data.. This extension has served to divert the activities of weather service away from theoretical phases of meteorology, to the solution of questions of immediate and pressing need.

It is yet too soon to assess the true value of free air data in meteorology. Some progress will be made in reaching a clearer conception of the physical processes concerned with weather changes; the improvement in weather forecasting will be slow until forecasters learn to interpret them in terms of coming weather. Their greatest and most immediate value, however, will be in the information they will convey to air pilots and the confidence that such information will beget.

Periodic fluctuations in climate.—Various investigators have sought to establish the existence of periodic fluctuations in climate; others have endeavored to determine what relations subsist between solar and terrestrial phenomena. A bibliography and a summary of the literature on this subject has been printed by the International Research Council, Paris, 1926. (First report of the Commission Appointed to Further the Study of Solar and Terrestrial Relationships.) In the solar-terrestrial group of investigators may be named the following: Abbot, Angot, Archibald, Alter, Angenheister, Arctowski, Baur, Bauer, Brückner, Blanford, Bigelow, Braak, Broun, Brooks (C. E. P.), Buchan, Chree, Clayton, Clough, Fritz, Hahn, Helland-Hansen and Nansen, Hildebrandsson, Huntington, Köppen, Liznar, Lockyer, Norman and W. J. S., MacDowell, Mielke, Newcomb, Nordman, Richter, Rizzo, Sekiguchi, Sverdrup, Unterweger, Walker, and others.

Helland-Hansen and Nansen [10] summarize the result of their extensive investigations as follows:

> The point of departure in these investigations was the wish to investigate more closely some of the yearly temperature variations in the North Atlantic Ocean. We have seen that such variations are present and that they are very considerable and extend over great regions in common. They can be ascribed in greater part to the action of the air pressure distribution, that is to say, the winds. In order to understand the occurrence and the nature of the variations, meteorological variations must therefore be closely studied. These can be understood only when the atmosphere as a whole is investigated, and we are therefore led to make a very wide investigation.
>
> Hitherto these extensive investigations have shown us that different groups of regions vary intact in a definite direction, while another group of regions varies in an opposite sense, and that again still other regions show transition phenomena, partly on account of phase displacements and partly on account of mixed relationships to the primary groups. All this gives us a variegated picture of the meteorological fluctuations, but out of this same variegated picture we find also by a proper analysis the influence of the variations in the solar activity which in all probability make themselves felt first in the higher layers of the atmosphere and thereby produce disturbances which again introduce changes in the lower layers. Such dynamic changes will take different courses in respect to the temperature, cloudiness, precipitation, etc., at different stations of the earth. But it seems possible by a thorough evaluation of available observational material to work out sure and general rules to cover the phenomena.

The present work is to be regarded only as an introduction to such more thorough investigations, and we must postpone a clarification of many of the questions raised here to later publications. Among them is the regulating action which the thermal condition of the ocean exercises upon the air circulation and the air temperature.

Meteorological effects of solar variability.—The results of researches on the relation between solar variability and terrestrial phenomena have been well summarized in the First Report to the International Research Council, by the commission appointed to further the study of solar and terrestrial relations (1926). The following, in brief, is the substance of the reports relating to meteorology, supplemented by references to other publications that seem pertinent, most of which are of a later date than that of the report.

Abbot [11] is convinced that the variability of the sun's output has been established, but that it is much less than he had once supposed.[12] He,[13] and also Clayton,[14] believes that a close relation exists between the value of the solar constant and meteorological conditions, and especially with respect to atmospheric pressure in certain regions. The argument for such a relationship is weakened by the fact that Clayton used "provisional values" of the solar constant, in his studies. The monthly mean values have since been corrected by amounts varying from +0.020 to −0.002 gr. cal. per minute per square centimeter, or by 1.14 per cent of the mean value of the solar constant.[15] The corresponding revised daily values have not yet been published.

Simpson [16] concludes that:

> Speaking broadly, one must say that the results up to the present time have been disappointing. Clear and definite relationships have not been found between sunspots and weather similar to those which have been found in the case of terrestrial magnetism. The eleven-year period is certainly recognizable in some meteorological factors, but in very few cases is the amplitude of any practical importance. The investigations into the day-to-day variations of solar radiation have been little more successful, although Mr. Clayton has exhibited many interesting and suggestive curves.

Walker [17] finds a correlation coefficient between sunspots and the annual temperature of India as given by the data of 47 years, as high as −0.5.

Brooks [18] has continued the investigations of Helland-Hansen and Nansen [10] to include the years 1914-1924, and has also summarized certain papers which deal with the general causes of the relationships between solar and terrestrial phenomena.

There is pretty general agreement that terrestrial temperatures, monthly and annual means considered, are a little higher than usual at times of minimum spottedness and a little lower than usual at times of maximum spottedness, and that the effect is more noticeable in the tropics than in temperature latitudes.

There seems not to be any definite evidence from the one hundred years instrumental observations available for examination that there are changes of a progressive or a cyclic character in the climate of any part of the earth.

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CHAPTER III

SOLAR RADIATION AND ITS RÔLE

Herbert H. Kimball

INTRODUCTION

The sun.—The sun is a gaseous body having a diameter of about 864,000 miles, which is more than 100 times the diameter of the earth. Its mass is more than 300,000 times that of the earth, and gravity at its surface is about 27.6 times the force of gravity at sea-level on the earth. In consequence, at the sun's center the pressure is enormous, and the temperature may reach 18,000,000° on the absolute centigrade scale, as compared with 6,000° at its surface. The distance between the sun and the earth varies from nearly 94.5 million miles early in July to about 91.3 million early in January.

Solar radiation.—The sun is practically the only external source of heat and light for the planets of the solar system. They intercept only an insignificant fraction of the total energy radiated from the sun (about 1/200,000,000). The intensity of this energy varies inversely with the square of the distance from the sun. It is therefore nearly 7 times as intense on the planet Mercury as on the Earth, and 27 times as intense on the Earth as on the planet Jupiter. Likewise, it is 1.07 times as intense when the Earth is in perihelion early in January, as when it is in aphelion early in July.

Radiant energy is scattered in passing through gases. It has been computed [1] that solar radiation in passing through the gases of the outer 5,000 miles of the solar sphere would be completely scattered. Therefore the sun radiates approximately as a black body at the temperature of this outer 5,000-mile layer of gas, or at about 6,000° absolute centigrade. Apparently, scattering and absorption in the solar gases cause the solar spectrum energy curve to differ somewhat from the spectral distribution of black-body radiation.

The exact nature of radiant energy is not certainly known. It exhibits the characteristics of wave motion transverse to its line of propagation. A fundamental law of wave motion is that in passing through the boundary surface between a rare and a dense medium the direction of motion is changed, unless the angle of incidence is zero. Also, the angular deviation increases with decrease in wave-length.

THE SOLAR SPECTRUM

The energy radiated by a hot body like the sun is polychromatic, but by passing it through a prism of glass or of some other transparent solid, like quartz, it will be arranged in the order of its wave-lengths. A visual examination shows the prismatic colors, violet at the end that has suffered the maximum deviation and red at the end least deviated, with the rest of the visible spectrum lying between. It is this part of the solar spectrum that is the source of natural light or daylight. By means of a heat-measuring device we will find that heat energy extends far beyond the red end of the visible spectrum, and this extension is known as the infra-red spectrum. Similarly, a heat-measuring device will show that heat energy extends beyond the violet to form the ultra-violet end of the spectrum. Sensitive photographic plates detect the presence of ultra-violet radiation to shorter wave-lengths than will the most delicate heat-measuring device.

Solar spectrum energy curves.—Langley [2] while at the Allegheny Observatory devised the spectrobolometer for measuring the relative intensity of heat energy in different parts of the spectrum. He and his successors with greatly improved apparatus [3] have obtained many bolographic records [4] from which have been determined solar spectrum energy curves at sea-level and also on high mountains with the sun at different zenith distances. It was found that in accordance with theory monochromatic radiation intensity may be expressed by the equation

$$I_\lambda = I_{0\lambda} a^m{}_\lambda \qquad (1)$$

where

I_λ = the measured intensity of a ray of wave-length λ,
$I_{0\lambda}$ = its intensity before depletion in the atmosphere,
a_λ = the atmospheric transmission coefficient, or the proportion of $I_{0\lambda}$ that would reach the surface of the earth with the sun in the zenith, and
m = the air mass, which is the relative length of the path of the solar rays through the atmosphere compared with the path when the sun is in the zenith, or approximately the secant of the sun's zenith distance.[5]

In the logarithmetic form (1) becomes $\log I_\lambda = \log I_{0\lambda} + m \log a_\lambda$, which is the equation of a straight line. Therefore, if logarithms of measured values of I_λ for a given place are plotted as ordinates against the corresponding values of m as abscissas, they will fall in a straight line, and extrapolation to zero atmosphere will give the corresponding logarithm of $I_{0\lambda}$.

In Figure 7, curve 1 is a reproduction of Abbot's[6] normal solar spectrum energy curve for zero atmosphere; curve 2 has been computed from 1 for the average atmospheric transmission at Calama, Chile;[7] curves 3-6, from 1 for the average atmospheric transmission at Washington,[8] and with the sun at zenith distances 0°, 60°, 75.°7 and 80.°7, respectively, the corresponding values of the air mass being 1.0, 2.0, 4.0, and 6.0.

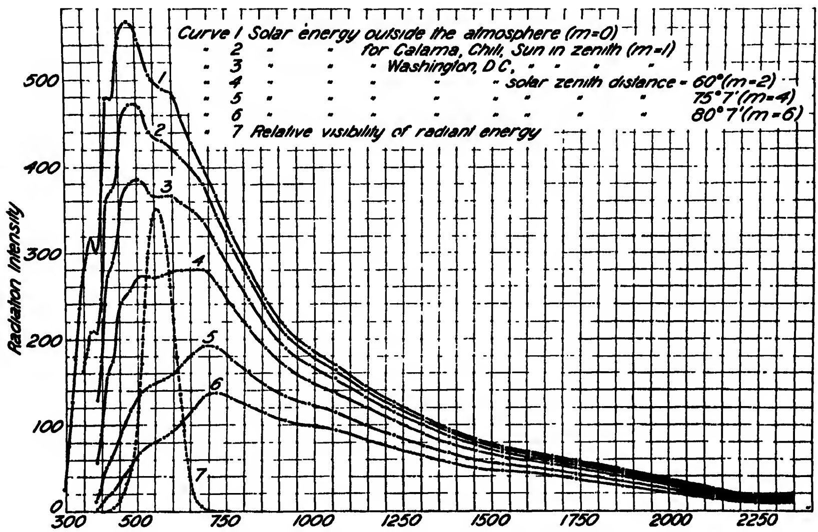

FIG. 7.—Solar normal spectrum energy curves, and curve of relative visibility of radiant energy.

ATMOSPHERIC DEPLETION OF SOLAR RADIATION

There are five principal causes of depletion of solar radiation by the earth's atmosphere in the absence of clouds, as follows:

1) Scattering by the gas molecules of pure dry air.

2) Absorption by the gases of dry air.

3) Scattering by the water vapor of the atmosphere.

4) Absorption by atmospheric water vapor.

5) Scattering and absorption by the liquid and solid particles, commonly called haze or dust, suspended in the atmosphere.

The curves of Figure 7 take into account the depletion by 1), 3) and 5), but do not include the selective absorptions of 2) and 4). The depletion by all five causes will now be considered.

Scattering by dry air.—The law of depletion of solar radiation by dry air was given long ago by Lord Rayleigh,[9] and more recently by King.[10]

Making use of King's equations, I have computed the transmission coefficients, $a_{a\lambda}$ for 38 values of λ between 0.3504μ and 2.442μ, for which Abbot has given what he considers the most reliable relative energy intensities, $I_{0\lambda}$, outside the atmosphere.[6] Then for the total spectrum we may write the equation

$$a'_{ma} = \frac{\Sigma(I_{0\lambda} a^m{}_{a\lambda})}{\Sigma I_{0\lambda}}. \qquad (2$$

Scattering by moist air.—Similarly, Fowle [11] has determined the transmission coefficient $a_{w\lambda}$, for a quantity of water vapor w, which, if precipitated, would form a layer of water 1 cm. thick. For w times this amount the transmission is $a^w{}_{w\lambda}$, and for air having a water vapor content of w centimeters of precipitable water

$$a'_{maw} = \frac{\Sigma I_{0\lambda}(a_{a\lambda} a^w{}_{w\lambda})^m}{\Sigma I_{0\lambda}}. \qquad (3)$$

When $w=0$ equation (3) becomes identical with (2).

At stations of the Astrophysical Observatory of the Smithsonian Institution the value of w is determined spectrobolometrically. At other stations it is necessary to use Hann's equation, $w=2.3e10^{\frac{-h}{22,000}}$, where e is the water vapor pressure in centimeters and h is the altitude above sea level in meters. Fowle [12] states that this equation can be relied upon only when normal values of e for a considerable period are available.

Values of atmospheric transmission with $m=0.526$, 1.0, 2.0, 3.0, and 4.0, and $w=0$, 0.5, 1.0, 2.0, 3.0, 4.0, 5.0, and 6.0, and also for $m=0.766$ with $w=0.0$, 0.42 and 1.28, have been computed from equation (3) by graphical methods, after applying a correction to both numerator and denominator of the second member for both ultra-violet and infra-red radiation beyond the wave-length limits to which reliable measurements have been made.[13] The results are given in Figure 8, curves 1 to 8, inclusive.

Absorption by atmospheric gases.—Figure 9 is a reproduction of a spectrobolometric energy curve obtained by Smithsonian observers on Mount Wilson, Calif., with a 60° ultra-violet crown-glass prism.[4, pl. 1] Note especially the oxygen band A, and the water vapor bands σ, ϕ, ψ, and Ω. There are also other water vapor bands, especially one at 0.81μ, and another at the extreme limit of the infra-red as shown in the figure. To compute the depletion represented by these bands I have utilized the curves given by Fowle,[11, Fig. 4] and my computed values of $I_{0\lambda}(a_{a\lambda} a^w{}_{w\lambda})^m$ for the values of λ covered by the bands.[11, p. 408] The computations have been made for the same values of m and w that were employed in Equation (3) to compute a'_{maw}. The results, expressed as a fractional part

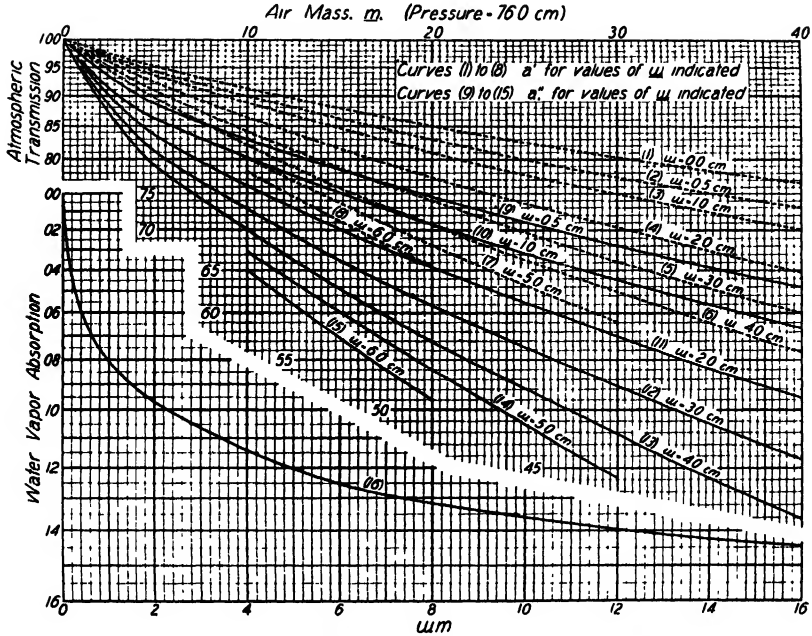

Fig. 8.—Atmospheric transmission of solar radiation through dust-free air.

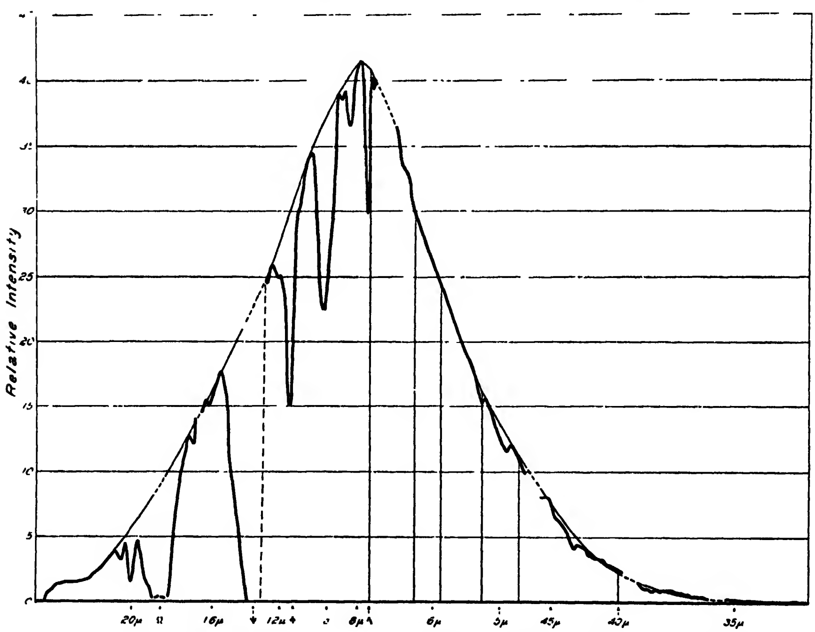

Fig. 9.—Bologram of solar spectrum of 60° ultra-violet crown-glass prism.

of $\Sigma I_{0\lambda}$, and increased by 0.005 to take account of band absorption by dry air,[11] have been subtracted from the corresponding values of a'_{maw} to obtain the values of a''_{maw} which are plotted in Figure 8, curves 9 to 15, inclusive.

The precipitable water in the path of the solar rays is equal to the product wm.[11, p. 398] On account of the high transmission coefficient of dry air at the wave lengths where water vapor bands occur, the depletion in these bands is practically the same for a given value of the product wm, within the limits of the respective values of the two factors here employed, and regardless of their respective values. Curve 16, Figure 8, which is the curve of best fit for computations with m having the values 1.0, 2.0, 3.0 and 4.0, and for each of these w having the values 0.5, 1.0, 2.0, 3.0, and 4.0 gives values that do not depart from the computed values by more than ± 0.001.

In Figure 8, the ordinates are plotted on the scale of their logarithms to the base 10. The abscissas have been numbered for an air pressure $P_0 = 76.0$ cm. With $P < P_0$, unit air mass falls on the scale of abscissas at P/P_0. Thus, if $P = 40.0$ cm., unit air mass, m', will fall at 0.526, $2.0m'$ at 1.052, etc. Likewise, for this new unit of air mass the water vapor content of each of the curves 1-15 inclusive, is $P/P_0 w$, instead of the value of w given.

*Depletion due to atmospheric dust.**—Curves 9 to 15, Figure 8, give atmospheric transmission coefficients, a''_{maw}, for dust-free moist air. It remains to determine the effect upon radiation intensities of smoke, haze, dust and liquid particles suspended in the atmosphere. For the sake of brevity this effect will be referred to as the dust transmission coefficient, a_d.

Let a_{0-1} represent the true atmospheric transmission, computed from the value of I_0 and a pyrheliometric measurement of solar radiation intensity with the sun in the zenith, I_1. Similarly, let a_{0-2} represent the

* While the manuscript of this chapter was in the hands of the publisher, a paper by A. Ångström was received entitled, On the atmospheric transmission of sun radiation, II. (Geografiska Annaler, 1930, H.2 O.3). In this paper Ångström maintains that the scattering by water vapor, $a_{w_\lambda}{}^{wm}$, measured by Fowle, is not molecular scattering, but in reality is the scattering by, the particles large compared with gas molecules, which are associated with the water vapor. Ångström would therefore include this depletion in the scattering by dust instead of in the scattering by dust-free air. Fowle (Astrophysical Journal, 42: 394) has indicated that at least part of the scattering which he has attributed to water vapor may be due to dust. If, after further consideration, Ångström's view is adopted, the "dust depletion" given in Table 3 must be increased by the scattering by water vapor indicated by curves 1 to 8, inclusive, Figure 8.

corresponding transmission when the pyrheliometric measurement is made with the sun 60° from the zenith ($m=2$), and let a''_{aw} and a''_{2aw} represent computed values of the transmission in an atmosphere free from dust, with a water vapor content, w, and the sun in the zenith and 60° from it, respectively. With the sun in the zenith the dust transmission coefficient $a_d=\frac{a_{0-1}}{a''_{aw}}$; with the sun 60° from the zenith, $a_{2d}=\frac{a_{0-2}}{a''_{2aw}}$; and for the sun at zenith distance $Z°$ (Air Mass $=m$)

$$a_{md}=\frac{a_{0-m}}{a''_{maw}}.$$

For Calama, Chile, during the warm and relatively moist months December to March, inclusive, with an average value of $w=1.28$ cm., $a_d=1.00$, or the effect of dust is negligible. During the remaining months, with an average value of $w=0.42$ cm., the average value of $a_d=0.973$.

Table 3 gives the monthly means of atmospheric transmission for dust-free air, as derived from Figure 8, and for dusty air as derived from pyrheliometric observations for three American stations, and Davos, Switzerland, in Europe. The difference gives the average depletion by dust each month, expressed as a fractional part of the value of the solar constant. This value of the depletion by atmospheric dust must not be confused with a_{md}.

For Washington, the average value of a_d varies from 0.875 to 0.923, and a_{2d} from 0.776 to 0.850, but Hand [14] has shown that on occasions when the smoke cloud over the city rises and passes over the pyrheliometric station at the American University the value of a_{md} is greatly reduced. Thus, on January 3, 1929, with $m=3.5$ and $w=1.29$, the intensity of solar radiation at normal incidence measured 0.22, or $a_{0-3.5}=0.11$. For dust-free air $a_{0-3.5}=0.59$ and $a_{3.5d}=0.11/0.59=0.19$. At the same time the intensity of the total solar radiation (direct+diffuse) received on a horizontal surface was decreased by 60 per cent. The dust count made during the passage of the cloud was no higher than is frequently observed in great industrial centers.

Furthermore, the average intensity of the total solar radiation received on a horizontal surface at the University Station, Chicago, Illinois, during the winter months December–February, is only 55 per cent of the intensity measured at Madison, Wisconsin, during the same months. Madison is 1° 18′ farther north than Chicago, and is by no means free from city smoke. Therefore, a conservative estimate of the average decrease in the intensity of solar radiation received on a horizontal sur-

face at Chicago by local smoke during the winter months is 50 per cent. When we remember that natural lighting diminishes in about the same proportion as the total solar radiation, the screening effect of city smoke on solar radiation assumes economic importance.

TABLE 3

DEPLETION OF SOLAR RADIATION BY ATMOSPHERIC DUST

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Washington, D. C.												
P (mm.)	754	753	752	751	750	750	750	751	753	753	753	754
W (mm.)	6.0	6.1	8.2	11.1	17.6	28.9	33.1	31.7	24.5	16.4	10.0	6.9
a_{0-1} (Pyr.)	....	....	0.71	0.71	0.68	0.65	0.63	0.64	0.69	0.72	....	....
Dust-free air	....	....	0.81	0.80	0.77	0.74	0.72	0.72	0.75	0.78	....	....
Dust depletion	....	....	0.10	0.09	0.09	0.09	0.09	0.08	0.06	0.06	....	....
a_{0-2} (Pyr.)	0.61	0.59	0.58	0.57	0.52	0.48	0.50	0.51	0.54	0.56	0.59	0.61
Dust-free air	0.74	0.74	0.71	0.70	0.67	0.61	0.60	0.60	0.64	0.67	0.71	0.73
Dust depletion	0.13	0.15	0.13	0.13	0.15	0.13	0.10	0.09	0.09	0.11	0.12	0.12
Madison, Wisconsin												
P (mm.)	736	737	736	735	735	736	736	737	737	737	737	737
W (mm.)	3.8	4.0	6.3	9.3	13.8	25.6	28.5	26.0	19.9	13.0	8.1	3.9
a_{0-1} (Pyr.)	....	....	....	0.74	0.72	0.71	0.68	0.69	0.71	0.72	....	....
Dust-free air	....	....	....	0.80	0.78	0.75	0.74	0.75	0.77	0.79	....	....
Dust depletion	....	....	....	0.06	0.06	0.04	0.06	0.06	0.06	0.07	....	....
a_{0-2} (Pyr.)	0.68	0.69	0.66	0.63	0.58	0.58	0.55	0.58	0.60	0.61	0.66	0.65
Dust-free air	0.76	0.76	0.74	0.72	0.69	0.68	0.62	0.63	0.66	0.70	0.73	0.76
Dust depletion	0.08	0.07	0.08	0.09	0.11	0.10	0.07	0.05	0.06	0.08	0.07	0.11
Lincoln, Nebraska												
P (mm.)	732	732	730	729	728	728	729	729	730	730	731	731
W (mm.)	5.1	5.7	7.7	11.6	16.1	28.2	30.7	29.5	21.0	11.8	8.7	6.0
a_{0-1} (Pyr.)	....	....	....	0.75	0.73	0.72	0.71	0.69	0.73	0.76	....	....
Dust-free air	....	....	....	0.80	0.78	0.74	0.74	0.73	0.76	0.80	....	....
Dust depletion	....	....	....	0.05	0.05	0.02	0.03	0.04	0.03	0.04	....	....
a_{0-2} (Pyr.)	0.70	0.69	0.64	0.62	0.59	0.58	0.57	0.56	0.61	0.61	0.68	0.68
Dust-free air	0.74	0.74	0.73	0.70	0.68	0.62	0.61	0.61	0.65	0.70	0.72	0.74
Dust depletion	0.04	0.05	0.09	0.08	0.09	0.04	0.04	0.05	0.04	0.09	0.04	0.06
Davos, Switzerland												
P (mm.)	625	626	627	628	629	630	630	629	628	627	626	625
W (mm.)	4.5	5.2	5.9	8.1	10.5	13.6	15.6	15.5	13.1	9.1	6.7	5.0
a_{0-1} (Pyr.)	....	....	0.81	0.81	0.79	0.78	0.75	0.78	0.79	....	....	....
Dust-free air	....	....	0.84	0.82	0.81	0.80	0.79	0.79	0.80	....	....	....
Dust depletion	....	....	0.03	0.01	0.02	0.02	0.04	0.01	0.01	....	....	....
a_{0-2} (Pyr.)	0.71	0.69	0.67	0.69	0.66	0.64	0.65	0.68	0.68	0.68	0.68	0.70
Dust-free air	0.77	0.76	0.76	0.74	0.72	0.70	0.69	0.69	0.71	0.73	0.75	0.76
Dust depletion	0.06	0.07	0.09	0.05	0.06	0.06	0.04	0.01	0.03	0.05	0.07	0.06

For any station the product $a_{md}a''_{maw}$ gives the true atmospheric transmission coefficient, a_m. Then $I_m = I'_0 a_m$, where I_m is the solar radiation intensity corresponding to air mass m, and I'_0 is the value of the solar constant adjusted to the relative distance of the earth from the sun at the time of year the intensity is desired. Since, however, a_{md} and a''_{maw}

can be determined for mean conditions only, the foregoing method of computing I_m should be employed only for weekly, monthly, or seasonal values.

MEASUREMENTS OF SOLAR RADIATION INTENSITY

Instruments.—For the measurement of the intensity of solar radiation as received at the surface of the earth various heat-measuring devices are available. Among the best-known are the following:

1) The Ångström compensation pyrheliometer,[15] which utilizes the thermopile, in connection with a sensitive galvanometer, to indicate when temperature equilibrium has been established between two blackened metal strips, one of which is exposed to solar radiation, and the other, which is shaded from the sun, is warmed by passing a measured current of electricity through it.

2) Thermoelectric pyrheliometers,[16] in which one set of junctions is warmed by exposure to solar radiation, while the other set is kept at the temperature of the air by shading it or by some other means. The electric current generated may be read by eye, or continuously recorded, as desired. Several different types of pyrheliometers employ this method of measurement.[17]

3) Instruments in which the rate of heating and cooling of a body, which is exposed to solar radiation and shaded from it during alternate intervals of time, is measured by a mercurial thermometer,[18] by an electrical resistance thermometer, as in the Marvin pyrheliometer;[19] or by a bi-metallic thermometer, as in the Michelson pyrheliometer.[20]

4) Photo-electric cells, which are sensitive to energy over only a restricted range of wave-lengths.[21]

Standard pyrheliometric scale.—The Ångström pyrheliometer was adopted as the standard instrument for solar radiation measurements at the Meteorological Conference at Innsbruck in 1906, and later in the same year at the Solar Physics Union at Oxford. Recognizing some structural defects in this instrument, Abbot, as early as 1907, had constructed a standard *water-flow* pyrheliometer, and later he constructed a standard *water-stir*, pyrheliometer.[22] The essentials of these instruments are a diaphragmed absorbing chamber inclosed in a calorimeter chamber which is filled with water. The pyrheliometer is calibrated by passing an electric current of known strength through a resistance coil in the absorbing chamber. It has been found that within 1 per cent of all the heat generated in the resistance coil is recovered in the calorimeter. Secondary instruments, such as Smithsonian Silver Disk pyrheliometers, have been standardized by comparison with these absolute instruments, when both

were exposed to solar radiation, and the constants of the secondary instruments thus obtained constitute the Smithsonian Revised Pyrheliometry of 1913.

Comparisons between the Ångström and Smithsonian instruments has shown that the latter gives results about 3.3 per cent the higher.[23] A. Ångström has conceded that 1.8 per cent of this difference may be attributed to certain features in the design of the Ångström instrument. The remaining 1.5 per cent must be attributed to difference in the two scales, the cause of which is yet to be determined. It is important to state on which of these scales pyrheliometric readings are given. In this chapter intensities are expressed in the Smithsonian Revised Pyrheliometry of 1913.

Radiation measurements.—These may be divided into three classes, as follows:

(a) Measurements of the intensity of solar radiation at normal incidence.

(b) Measurements of the intensity of the total solar radiation received on a horizontal surface directly from the sun and diffusely from the sky.

(c) Measurements of the intensity of solar radiation received from a limited section of the spectrum, as the ultra-violet or the visible.

Measurements of class (a).—Solar radiation measurements at normal incidence are chiefly interesting in studies of atmospheric transmission, which diminishes with increase of atmospheric pressure, and with the water vapor and dust content of the atmosphere, as has already been shown. This is further illustrated by the atmospheric transmission coefficients at different altitudes given in Figure 10.

Measurements of class (b).—Of the solar radiation that is diffusely scattered in its passage through the atmosphere, a considerable portion eventually reaches the surface of the earth. Since it comes from every point of the hemispherical vault of the sky, it is most conveniently measured by a pyrheliometer having its receiving surface horizontally exposed. Furthermore, since it must be exposed to the whole sky, and at the same time protected from wind and rain, it is usually placed under a cover of glass that transmits the short-wave solar radiation, but which completely shuts off energy of long wave-lengths radiated by bodies at low temperatures, such as the earth and its atmosphere. A pyrheliometer thus exposed will therefore measure the intensity of the vertical component of the radiation received directly from the sun and diffusely from the sky. This intensity has been referred to as "The fundamental basis of the science of meteorology."[24] Published records of this intensity are available from only about 28 stations.

In Figure 11, curve *BB* is a record of the intensity of the total solar radiation received on a horizontal surface from the sun and sky; curve *CC* represents that part of the radiation that is received from the sky; curve *AA* the intensity of direct solar radiation at normal incidence.

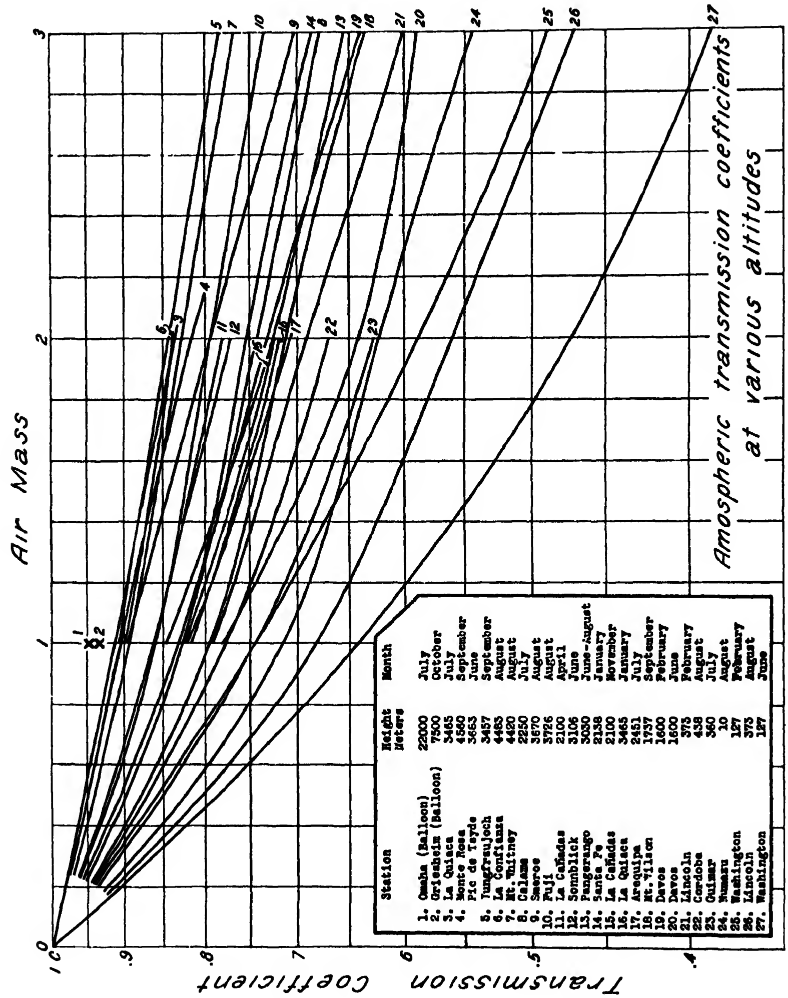

FIG. 10.—Atmospheric transmission coefficients at different altitudes.

The intensity scale is an arbitrary one. To obtain intensities in gram-calories per minute per square centimeter, the ordinates of curve *AA* must be multiplied by 0.064, and the ordinates of *BB* and *CC* by 0.073.

A list of stations at which measurements of classes 1) and 2) as designated above have been made was prepared and published by me

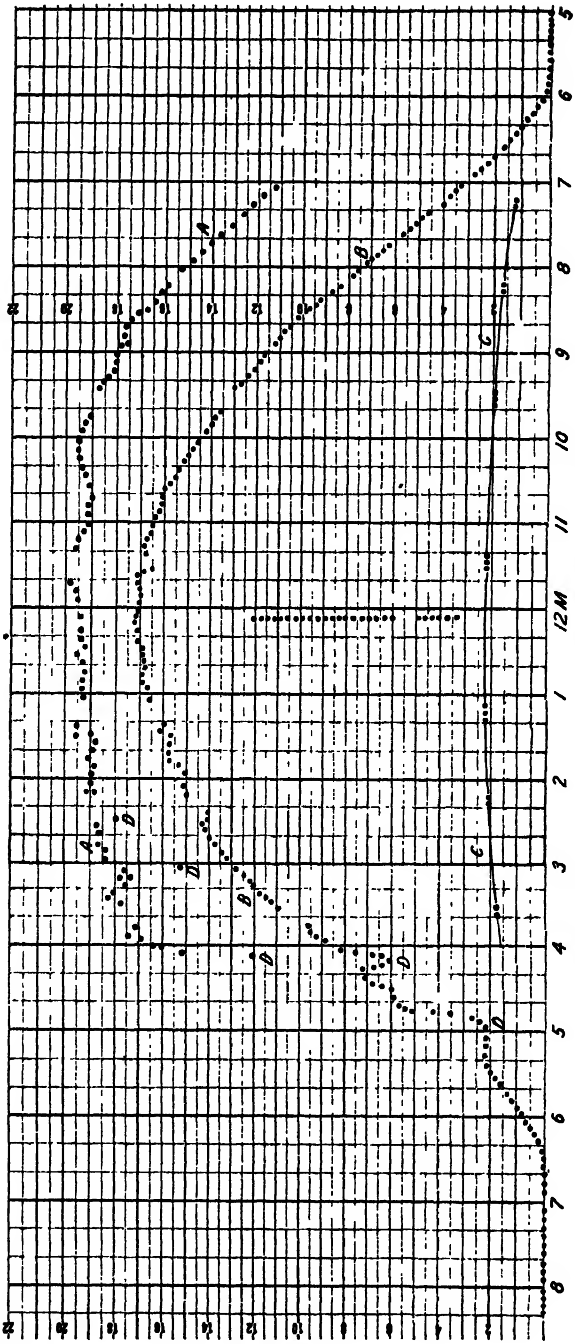

FIG. 11.—Record from the U. S. Weather Bureau thermoelectric pyrheliometer.

in 1927.[25] A second paper supplies data to the end of 1929.[26] In this latter publication, Table 3 gives monthly means and Table 4 weekly means of the daily totals of radiation of class (*b*) for 23 widely scattered stations, and Table 5 gives annual totals for 13 stations. The excess in the annual totals at low latitude as compared with high latitude stations is clearly shown.

Cloudiness is an important factor in determining the daily or annual totals of solar radiation. Representing by Q_0 the total radiation received on a cloudless day, by Q_s the amount received on a day with clouds, and by S the ratio $\frac{\text{recorded duration of sunshine}}{\text{possible duration of sunshine}}$ Ångström found that for Stockholm $Q_s = Q_0(0.235 + 0.765\,S)$. For Washington, I have found that $Q_s = Q_0(0.22 + 0.78\,S)$, which is very close agreement with Ångström.[27] In accordance with these equations the total radiation on a day with the sky completely covered with clouds averages from one-fifth to one-fourth as much as it does on a day with no clouds.

There is a marked diurnal variation in the total radiation received on a horizontal surface, as is shown by the curves of Figure 11. Outside the tropics the annual variation is strong, and there is also a variation with latitude. Both variations are shown by the isopleths of Figures 12-17, which give daily totals of solar radiation received over the oceans,[28] expressed in gram-calories per cm.² The computations are based on average daily cloudiness and water vapor content of the atmosphere as derived from various sources,[29, 30, 31] and such pyrheliometric measurements as are available for determining the dust transmission coefficient, a_{md}. Figure 8 has been utilized in making the computations for dust-free air.

Measurements of class (*c*).—For measurements of the intensity of solar radiation in limited sections of the spectrum, reference should be made especially to the work of Dorno at Davos, Switzerland,[32, 33] and Gorczyński in North Africa and elsewhere.[34]

Gorczyński isolated different sections of the spectrum (blue, yellow, red and infra-red) by means of light filters. His measurements show, what would be expected from theory, that the red and infra-red in the solar spectrum increase in their proportion to the total spectrum with decrease in solar altitude, and also with increase in latitude.

Dorno has also used light filters for studies of the variations in the intensity of different sections of the visible spectrum, and, in connection with photo-electric cells, for the measurement of the intensity of the ultra-violet in solar radiation. Especial attention has been given by him to measurements of this section of the solar spectrum, and to a study of its diurnal, annual and secular variations.[33] The diurnal and annual

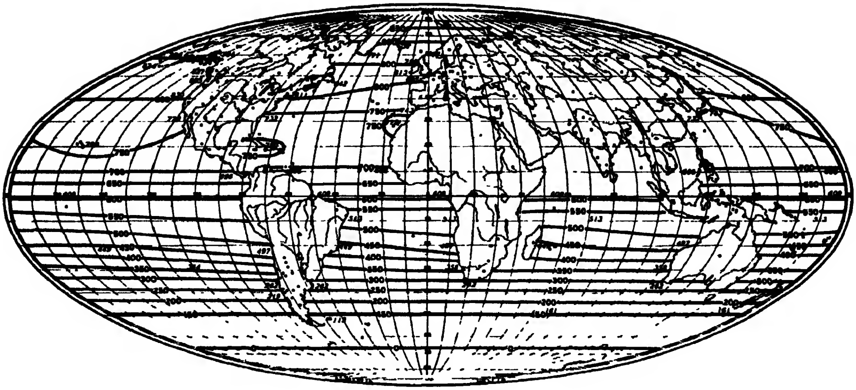

FIG. 12.—Isopleths of total solar radiation on June 21, with cloudless sky.

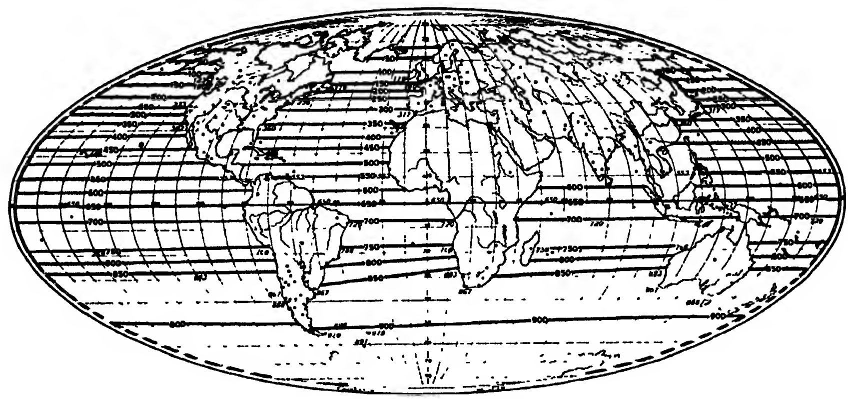

FIG. 13.—Isopleths of total solar radiation on December 21, with cloudless sky.

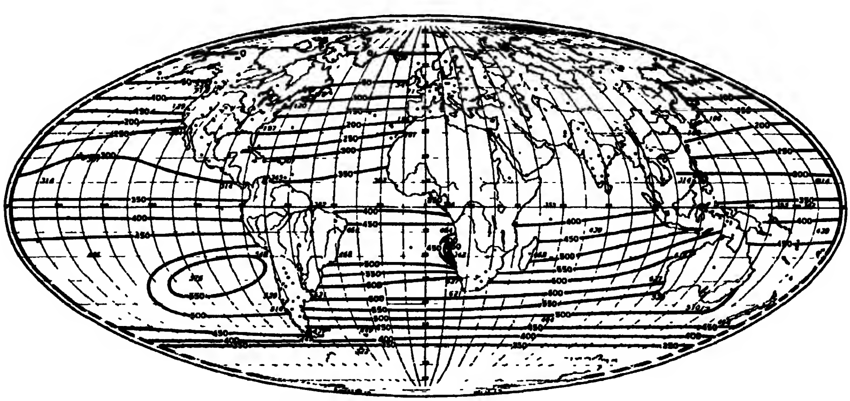

FIG. 14.—Isopleths of total solar radiation on December 21, with average cloudiness.

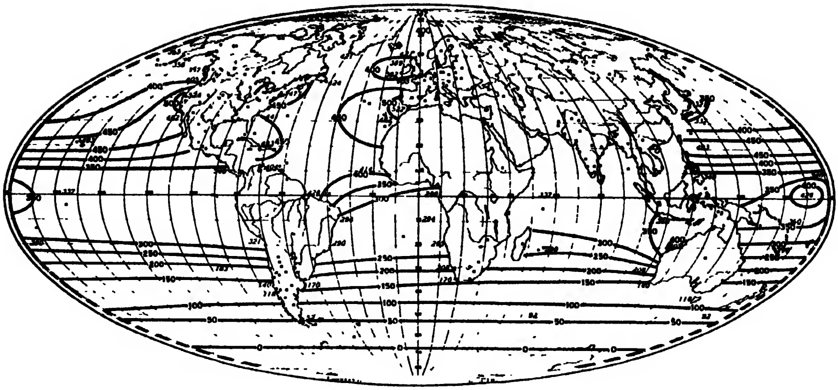

Fig. 15.—Isopleths of total solar radiation on June 21, with average cloudiness.

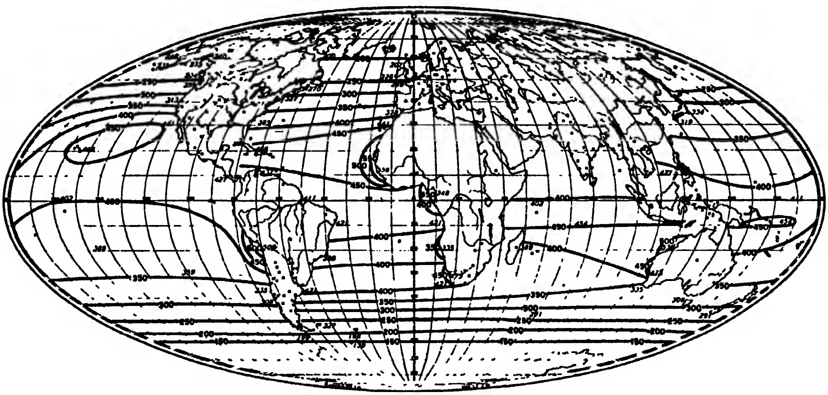

Fig. 16.—Isopleths of total solar radiation on March 21, with average cloudiness.

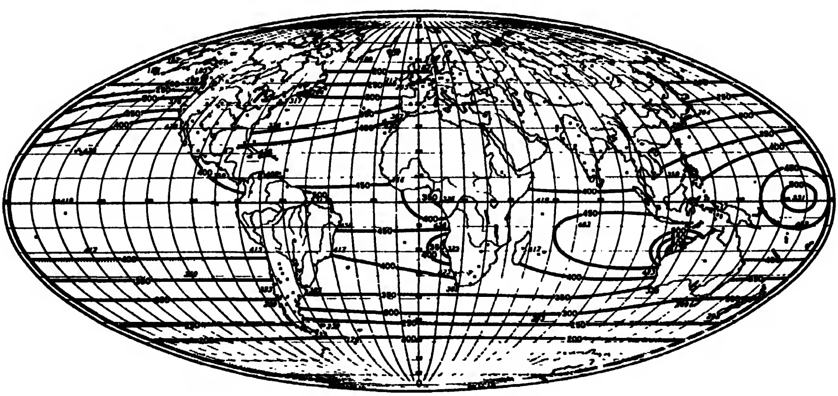

Fig. 17.—Isopleths of total solar radiation on September 21, with average cloudiness.

variations are shown in a table which gives, in relative measures, the mean intensity of the ultra-violet at different hours of the day in different months. The maximum occurs at mid-day in August, and the minimum for mid-day occurs in December. Another table gives the shortest wavelength measured at different hours of each month. The shortest measured at any time is 293.9$\mu\mu$ between noon and 1 P. M. in April. The shortest measured in December is 306.2$\mu\mu$ between the same hours. Between 5 A. M. and 6 A. M. in August the shortest wave-length measured is 320.2$\mu\mu$. It is apparent that the energy between wave-lengths 302 and 298$\mu\mu$, which is considered most effective in biological reactions, is found only with high sun.

Interesting comparisons have been made by Pettit [35] at Mount Wilson, California, on the relative intensity of the ultra-violet and the green regions of the solar spectrum. He concludes that the intensities of the ultra-violet and of the total solar spectrum vary in the same direction, and that the percentage change is much the greater in the ultra-violet.

DETERMINATIONS OF THE SOLAR CONSTANT OF RADIATION

For many years the Astrophysical Observatory of the Smithsonian Institution has employed the following method for determining the intensity of solar radiation before depletion by the Earth's atmosphere.

Long method.[36]—

1) From spectrobolographs similar to Figure 9 a series of solar spectrum energy curves corresponding to curves 3-6, Figure 7, are obtained.

2) By extrapolation, the extra-terrestrial solar spectrum energy curve is formed, and corrected for ultra-violet and infra-red radiation not measured by the bolometer. A graphical integration gives $\Sigma I'_{0\lambda}$.

3) Similarly, a terrestrial solar energy curve is corrected and integrated, taking account also of depletion by atmospheric absorption, especially in the great water vapor bands shown in Figure 9. The integration gives $\Sigma I_{m\lambda}$, and corresponding to it is a pyrheliometric measurement of the solar radiation intensity, I_m.

Then

$$I'_0 = I_m \frac{\Sigma I_{0\lambda}}{\Sigma I_{m\lambda}},$$

and the solar constant, $I_0 = I'_0 R^2$, where R is the distance of the earth from the sun in terms of its mean distance.

Short method.[37]—From one bologram to determine the water vapor content of the atmosphere, and a measurement by the pyranometer of the brightness of the sky about the sun, a function F is derived, which is definitely related to the atmospheric transmission coefficient, a_λ, for different wave-lengths of radiation. By this method, therefore, it is pos

sible to determine from one observation the values of a_λ that formerly called for the making of several bolograms, requiring perhaps 2 to 4 hours of time, during which time-interval the atmospheric transmission may have changed, giving an erroneous value of the solar constant.

On the other hand, it has been found necessary to obtain observations over several months at a given station before the relation between the function, F, and atmospheric transmission, a_λ, is definitely fixed.[38]

VARIATIONS IN SOLAR RADIATION INTENSITIES

The solar constant.—The solar constant determinations by the Smithsonian Institution cover the period 1902 to the present time. The progressive decrease in day-to-day variations in value is testimony to improvement in both instrumental equipment and methods of observation and reduction.[38, 39] Although the final revision of all solar constant determinations has not yet been published, the mean value appears to be close to 1.94 gram-calories per minute per square centimeter. During the period August 1918–February 1930, the highest monthly mean value is 1.969 in September 1921, and the lowest, 1.912 in July 1922, a range of 3 per cent.

Abbot[40] found periods of 2.14, 1.28 and 0.91 years in solar constant values for the years 1920-1926. With these periods as a basis he predicted "An appreciable maximum of solar constant values about May 1928, and very decided minimum about November 1928." The highest mean of three consecutive values in 1928 was 1.957 in May and the lowest corresponding mean was 1.911 in September, giving a range of 2.4 per cent.

Intensity at normal incidence.—A summary of monthly means of solar radiation measurements made at widely separated points and extending over considerable periods[41] is presented graphically in Figure 18. It will be seen that in the period 1883-1923 there have been marked depressions in the monthly means, following, respectively, the three great explosive eruptions of Krakatoa in 1883, Pelée, Santa Maria and Calima in 1902, and Katmai in 1912. The secondary depressions in 1888-89 and 1890-91 also followed periods of explosive volcanic eruptions.

From 1914 to the present time the monthly means have shown little variation. Their increased smoothness over that of years prior to 1897 is partly due to the increased number of stations at which measurements are now made.

Total radiation received on a horizontal surface.—Since fine dust such as is thrown into the high layers of the atmosphere by an explosive volcanic eruption depletes the incoming solar radiation principally through scattering, the depletion of the total (direct+diffuse) received on a

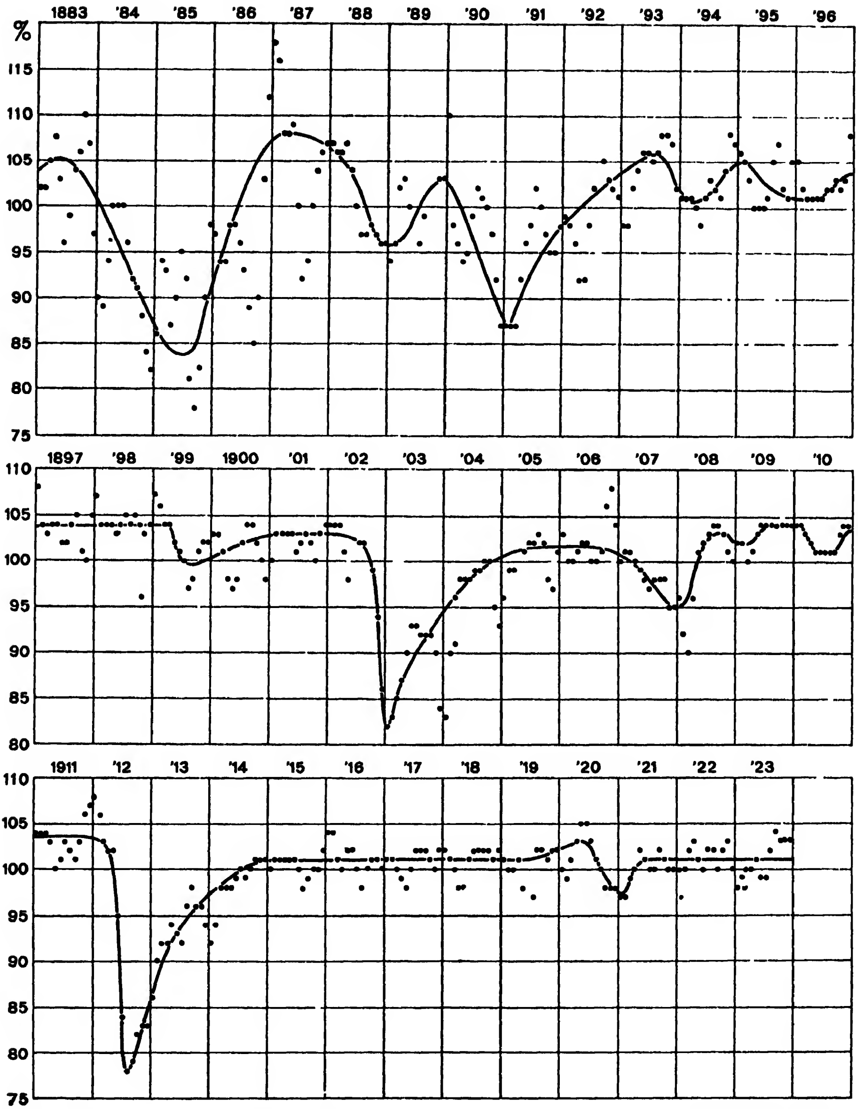

FIG. 18.—Monthly averages of solar radiation intensity at the Earth's surface expressed as a percentage of monthly normals.

horizontal surface will be much less than is the case with direct solar radiation. In 1912, at Mount Weather, Virginia, the total solar radiation received on a horizontal surface in the period September--December, on clear days averaged about 4 per cent less than in the corresponding period of 1913; but this is not larger than annual deficiencies that have occurred in years when volcanic dust was absent, and which appear to be due to unusual cloudiness.

From the foregoing we must conclude that variations in atmospheric depletion cause greater variations in the total receipt of solar radiation at the surface of the earth than do any changes in the solar constant that have yet been established.

In connection with the construction of Figure 8 it was necessary to determine the intensity of solar radiation in different parts of the spectrum both for dry air and for air containing different amounts of water vapor. The determinations were made for sea level conditions, $B=760$ mm., and also for the height of Calama, Chile, $B=582$ mm., $m'=\frac{582}{760}=0.766$. Having determined these intensities, it is a simple matter to compute the relative energy in the solar spectrum between different wave-length limits. For dust-free air at an altitude of 2250 meters the results are given in Table 4, and for dust-free air at sea-level, in Table 5 (A). In Table 4 the second column also gives the distribution of energy outside the atmosphere ($m'=0$).

TABLE 4

PERCENTAGE OF TOTAL ENERGY IN DIFFERENT PARTS OF SOLAR SPECTRUM OUTSIDE THE ATMOSPHERE ($m=0$) AND WITH A DUST-FREE ATMOSPHERE AT THE ALTITUDE OF CALAMA, CHILE ($m'=\frac{582}{760}=0.766$); ALSO DEPLETION BY ATMOSPHERIC DUST (DEPLETION IN TOTAL SPECTRUM = 10 PER CENT OF SOLAR CONSTANT).

	Energy distribution				Depletion by dust		
Air mass (m')	0	0.766	0.766	0.766	0.766	0.766	0.766
Water-vapor content of atmosphere, cm.	0	0	0.42	1.28	0	0.42	1.28
Place in spectrum	%	%	%	%	%	%	%
Below 0.346μ	3.1	1.8	1.7	1.5	29.1	30.4	31 9
0.346–0.405μ	5.0	3.8	4.0	4.1	23.6	24.8	26.0
0.405–0.704μ	40.1	39.5	42.3	43.6	14.9	15.6	16.4
Above 0.704μ	51.8	54.9	52.0	50.8	7.0	7.1	7.1
Total	100	100	100	100	....	....	....
Percentage of solar constant	100	93.5	86.0	81.2	10.0	10.0	10.0

Angström[42] has recently shown that the depletion of solar radiation by atmospheric dust may be expressed by the equation

$$\gamma=\frac{\beta}{\lambda^{a}}$$

He has also found that under normal conditions the value of a departs but little from 1.28, although following the eruption of Katmai Volcano in 1912 its value was reduced about one-half. I have applied this equation to the intensities of solar radiation at different wave-lengths after depletion by dust-free air, including the absorption by ozone, water vapor and other gases of the atmosphere. Table 4 and Table 5(B) give the resulting percentages of depletion by atmospheric dust, computed for a depletion in the total spectrum amounting to 10 per cent of the solar constant. For any other percentage of depletion, as x, the percentage given in the tables must be multiplied by $\frac{x}{10}$. At Davos, in April, August and September, $m'=\frac{616}{760}=0.811$, the depletion by atmospheric dust amounts to only one per cent, and interpolation between Table 4 and Table 5(B) indicates that the depletion in the different spectral regions, beginning with the shortest wave-length, would be about 3.1, 2.5, 1.6, and 0.7 per cent, respectively.

On the other hand, at Washington in September, with the sun 60° from the zenith ($m=2$), an atmospheric water vapor content of 2.45 cm., and a depletion by atmospheric dust of 9 per cent, the percentage depletions obtained from Table 5(B) for the different spectral regions are, respectively 38.2, 31.0, 18.1, and 9.3.

The corresponding intensities in these spectral bands may be obtained from Tables 5 (A) and 5 (B) as follows:

Intensity below 0.346μ

$$=0.002(1.94\times0.637)(1-0.382)=0.0015$$

Intensity between 0.346μ and 0.405μ

$$=0.024(1.94\times0.637)(1-0.310)=0.0205$$

Intensity between 0.405μ and 0.704μ

$$=0.438(1.94\times0.637)(1-0.181)=0.4433$$

Intensity above 0.704μ

$$=0.536(1.94\times0.637)(1-0.093)=0.6008$$

Intensity in total solar spectrum with earth at mean solar distance $=1.0663$

Intensity reduced to mean solar distance for September $=\frac{1.0663}{R^2}=1.055$ gram calories per minute per square centimeter, or practically the same as in Table 3 $\left[(1.94\times0.54)\frac{1}{R^2}=1.048\right]$.

It must be understood that the above computation of intensities in different parts of the spectrum are for average conditions only, and take no account of the effect of variations in the ozone content of the atmosphere with both time and place, or of possible variations in the distribution of energy in the spectrum of solar radiation before it enters the

TABLE 5

(A) Percentage of Total Energy in Different Parts of Solar Spectrum After Depletion by Dust-Free Air

Solar zenith distance		0°				60°				75.7°			
Air mass.	0	1				2				4			
Water-vapor content of atmosphere (cm.) ..	....	0.0	1.0	2.0	3.0	0.0	1.0	2.0	3.0	0.0	1.0	2.0	3.0
Place in spectrum	%	%	%	%	%	%	%	%	%	%	%	%	%
Below .346μ	3.1	1.3	1.2	1.0	0.7	0.6	0.5	0.3	0.1	0.2	0.1	0.06	0.02
.346-.405μ	5.0	3.5	3.8	3.7	3.6	2.4	2.7	2.5	2.3	1.1	1.2	1.07	1.00
.405-.704μ	40.1	39.3	43.3	44.2	44.9	37.6	42.3	43.4	44.2	34.0	39.7	40.0	40.90
Above .704μ	51.8	55.9	51.7	51.1	50.8	59.4	54.5	53.8	53.4	64.7	59.0	58.5	58.10
Total spectrum ..	100	100	100	100	100	100	100	100	100	100	100	100	100
Percentage of solar constant..	100	90.4	80.3	76.8	73.1	84.2	69.2	65.9	61.1	75.8	58.8	51.7	45.8

(B) Depletion of Energy by Atmospheric Dust (Total Depletion = 10% of Solar Constant)

Below .346μ	.....	30.3	32.7	34.4	35.7	32.4	40.9	41.8	43.3	41.5	48.7	56.4	63.5
.346-.405μ	......	24.6	26.7	27.9	29.0	26.5	31.8	33.7	34.9	33.2	39.1	45.2	50.2
.405-.704μ	......	15.5	16.8	17.5	18.3	18.1	19.7	20.9	21.6	20.4	25.0	27.5	30.5
Above .704μ	......	6.4	7.3	7.8	8.2	6.9	9.0	9.5	11.2	8.7	11.2	13.2	15.2

(C) Total Depletion = 10% of Intensity of Dust-Free Air

Below .346μ	......	27.4	26.3	26.2	26.1	27.3	29.2	27.5	26.5	31.5	28.6	29.2	29.1
.346-.405μ	......	22.2	21.4	21.3	21.2	22.3	22.7	22.2	21.3	25.2	23.0	23.4	23.0
.405-.704μ	......	14.0	13.5	13.4	13.4	15.2	14.0	13.8	13.2	15.5	14.7	14.2	14.0
Above .704μ	......	5.8	5.9	6.0	6.0	5.8	6.2	6.3	6.8	6.6	6.6	6.8	7.0

earth's atmosphere. Therefore they cannot take the place of careful measurements, and only serve to indicate approximately how much energy is to be expected in different spectral regions under different atmospheric conditions. The variations in the ozone content of the atmosphere must exert a marked effect upon the amount of solar energy of wave-lengths below 0.346μ that reaches the surface of the earth.

Daylight intensity.—Curve 7 in Figure 7 gives the relative visibility of radiant energy of different wave-lengths. It will be noted that most of the visible rays lie between 400 and $700\mu\mu$. Also, since the depletion by scattering is so much greater at the blue end than at the red end of the spectrum, the skylight, which is made up of diffusely scattered sunlight, is blue in color.

The photometer is used to measure the brightness of the sky and the intensity of natural or daylight illumination. In the absence of clouds, the brightness and color of the sky vary with angular distance from the sun on both the azimuth and altitude circles. Figure 19 is a stereographic projection of the half of the sky on one side of the sun's vertical. The sky brightness should be symmetrical on each side of this vertical circle. The position of the sun is indicated by x, 40° above the horizon.

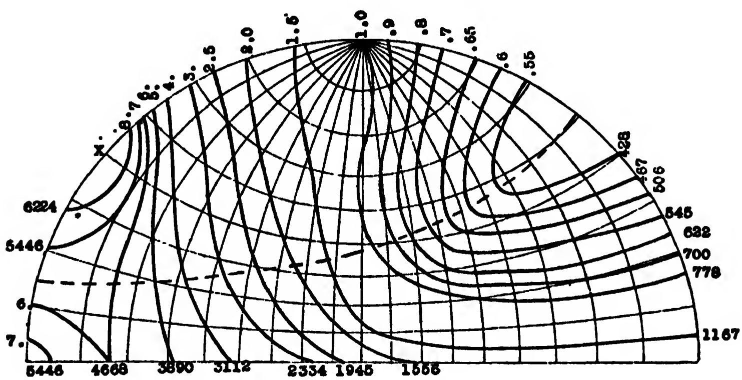

FIG. 19.—Distribution of sky brightness with a cloudless sky and the sun 40° above the horizon.

The figures at the top give the brightness relative to zenith brightness; the figures at the opposite ends of the lines of equal brightness give the brightness in millilamberts. The darkest point in the sky is nearly 90° from the sun on the same vertical circle, and this is also the bluest point. Brightness increases towards the horizon, and also with proximity to the sun. Similarly, Figure 20 shows the distribution in brightness when the sky is covered with dense clouds. The sky is brightest near the zenith and gradually grows darker towards the horizon.

The brightness of the sky as a whole increases with increase in the altitude of the sun. For this reason a cloudy day in winter is darker than a cloudy day in summer, and other things being equal the daylight intensity increases with decrease in latitude.

Figure 21 gives the illumination on a horizontal surface, with a cloudy sky of average density, for each hour of the day throughout the year at

latitude 42° North. This figure is important, since the illumination with such a sky is considered the standard daylight intensity for which the fenestration of buildings to be illuminated by natural light must be planned. Latitude 42° North was selected for this illustration because so many important industrial cities are near this parallel, such as Boston, New York, Pittsburgh, and the cities on the south shores of the Great Lakes. The illumination increases at the rate of about 17 foot-candles for each degree of increase in the sun's altitude. At noon, therefore, cities at latitude 40° North should have illumination intensities about 34 foot-candles higher than are given in Figure 21, which is a small percentage increase in summer, but becomes of importance in midwinter.

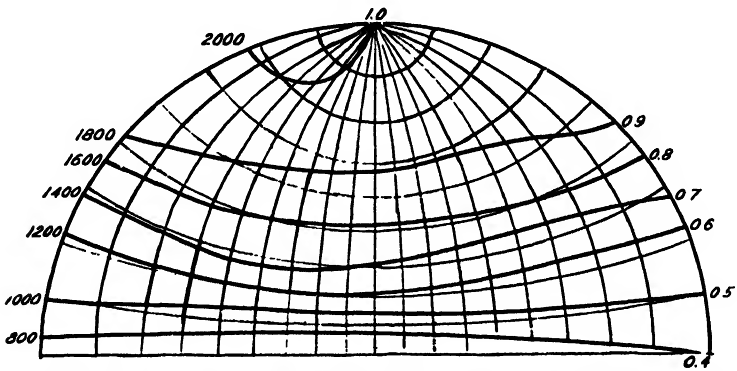

FIG. 20.—Distribution of sky brightness with the sky covered with dense clouds, and the sun 40° above the horizon.

At latitude 44° North, the illumination would be less at noon by about 34 foot-candles than is given in Figure 21.

With an unobstructed horizon, the illumination from a cloudy sky on a vertical surface averages about 14 per cent less than on a horizontal surface. Unobstructed horizons are so rare in industrial plants that the above figures have but little practical significance.

On clear days in midsummer in temperate latitudes, natural illumination on a horizontal surface may reach 10,000 foot-candles.

The color temperature of daylight.—Natural light, or daylight, is made up of the light received directly from the sun plus that received diffusely from the sky. Referring to Figure 7, curve 1, the distribution of solar energy outside the atmosphere approximates that of black-body radiation at a temperature of 6,000° absolute on the Centigrade scale. That is to say, its color temperature is 6,000° K. With passage of the solar rays through the atmosphere, the maximum of the energy curve

shifts towards the red end of the spectrum, the shift increasing with increase in the value of the air mass m. On the other hand, the proportion of the total daylight that is received diffusely from the sky increases with approach of the sun to the horizon,[43] and at the same time the proportion of blue and violet light in skylight increases. Priest's measurements [44] indicate that with the sun on or near the horizon the color temperature of skylight may approach 25,000° K., and that at mid-day it may approach 10,000° K. From this it results that there is but little departure from 6,000° K. in the color temperature of total daylight at

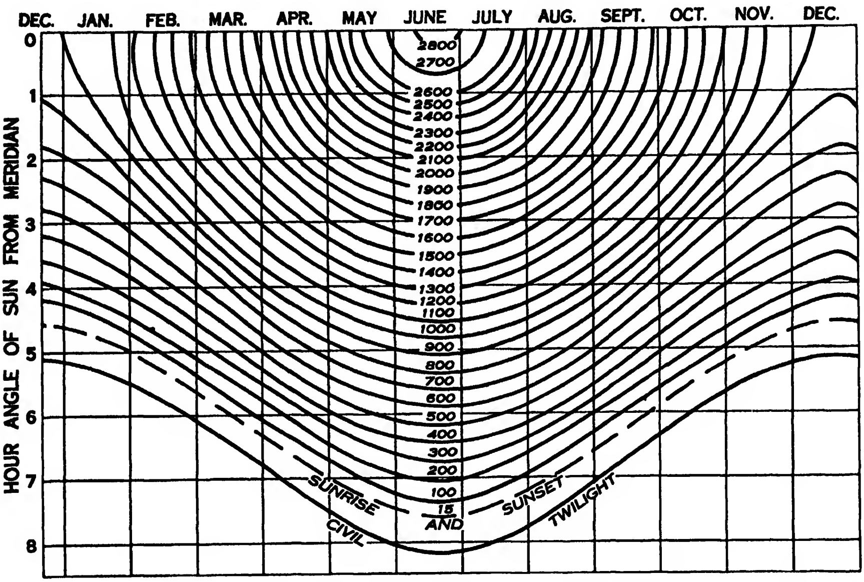

FIG. 21.—Illumination on a horizontal surface at latitude 42° N., with the sky covered with clouds of average thickness.

sea-level with the time of day or season of the year, except when the sun is near the horizon. There is an increase in its color temperature with altitude above sea-level.

THERMODYNAMIC EFFECTS OF SOLAR RADIATION

The thermodynamic effects of the periodical variations in the intensity of solar radiation as received at the surface of the earth are well known. In general, after sunrise on each clear day the air temperature increases, rapidly at first, and then more slowly, until a maximum is reached from two to four hours after noon. The air then begins to cool, the cooling becoming most rapid soon after sunset. The minimum air temperature is reached just before sunrise.

If the surface heating is excessive and the air sufficiently moist, thunderstorms, often of great violence, may result. A cloudy sky decreases the intensity of the incoming heat energy and diminishes the rise of temperature during the day.

In response to the annual period in the total solar radiation received, the daily means of temperature also show an annual period; but the maximum is reached not at the time of the summer solstice, but about a month later, and the minimum temperature about a month after the winter solstice.

Solar radiation and air temperature.—Ångström [45] has shown that the relation between the intensity of solar radiation received at the surface of the earth and surface air temperature may be expressed mathematically. But in order to do this he first considers what becomes of the radiation received. A part, after absorption, is reradiated to the atmosphere as low-temperature or long-wave radiation.* The atmosphere, in turn, radiates to the earth. The net result is a loss of heat to the atmosphere throughout the 24 hours. Another part of the solar radiation received is reflected back to the sky, the amount depending upon the character of the surface upon which it is received. Still another part is expended in evaporating moisture from the surface of the earth.

The net loss of heat by radiation to the atmosphere may be measured, provided the earth radiates as a black body. An instrument that has been used for such measurements employs the principle of the Ångström electrical compensation pyrheliometer.[46] The Melikeron [47] has also been used successfully, and in connection with a recording microammeter has given continuous records of the intensity of outgoing radiation during the night.

Measurements of the albedo of different kinds of surfaces [48, 49, 50] and determinations of the rate of evaporation from the surface of the earth [51] have been employed to determine losses of radiation by reflection and evaporation.

Representing by Q_m the monthly or weekly means of solar radiation received on a horizontal surface, and by R_m the corresponding means for the outgoing radiation, if from $Q_m - R_m$ we deduct the corresponding losses due to reflection and evaporation, there remains a quantity which may be represented by Q_T, and which Ångström designates the *temperature effective energy,* or that part of the radiant energy received at the surface of the earth that is available for heating the atmosphere.

* This low-temperature radiation must not be confused with solar radiation received diffusely from the sky.

Ångström further shows that the monthly means of solar radiation for Stockholm may be expressed by the Fourier series

$Q_m = 6{,}100 + 6{,}130 \sin (296.5° + x) + 590 \sin (170° + 2x) + 440 \sin (109° + 3x)$, and the monthly means of the *temperature effective energy* by

$Q_T = -270 + 3{,}740 \sin (295.2° + x) + 800 \sin (185.2° + 2x)$.

Also, the mean air temperature may be expressed by

$$T = 5.9° + 9.7° \sin (262.2° + x) + 1.0 \sin (90.0° + 2x) + 0.2 \sin (288.5° + 3x)$$

In the above $x = 0°$ on January 15, and Q_m and Q_T are expressed in gram-calories per cm².

For Washington, I find for the daily totals of radiation, expressed in the above units, and for $x = 0$ on July 5,

$$Q_m = 335.4 + 171.4 \cos (x + 13.1°) - 20.0 \cos (2x + 3.4°);$$
$$Q_T = 32.2 + 106.4 \cos (x + 10.0°) - 22.3 \cos (2x - 33.0°);$$

and for the mean daily temperature,

$$T = 12.8° + 12.3° \cos (x - 15.5°) + 0.3° (2x + 45.9°).$$

Ångström [45, 52] finds a close relation existing between Q and T, and he is therefore able to show what would be the effect of a change in the radiation receipt due to a change in the average cloudiness. For example, the total absence of clouds would give an annual temperature 1.8° higher and an annual amplitude 6.3° greater than at present. That is to say, the summers would be warmer and the winters colder than now. Also, with the present percentage of possible sunshine of 39 throughout the year instead of the existing cloudy winters and comparatively clear summers, the average annual temperature would be 1.2° lower and the annual amplitude 1.1° less than now.

Ångström also computes [45] that an increase in the value of the solar constant of 3 per cent from sunspot minimum to sunspot maximum would cause the annual temperature at Stockholm to be 0.4° higher at the latter epoch.

In Figure 22, curve 5 gives the average daily totals of radiation received at Washington throughout the year, and curve 7 gives the corresponding daily averages of temperature. Curve 3 gives the average daily totals of radiation that would be received at Washington with a continuously clear sky. I have computed that the mean air temperature corresponding to curve 3 may be represented closely by

$$T_{co} = 19.4° + 16.7° \cos (x - 15.5°)$$

which gives a mean annual temperature 6.6° higher and an annual range

8.8° greater than at present; that is, it would be about 3° warmer in winter and 11° warmer in summer than at present.

In the absence of snow during the winter I have computed for the temperature effective radiation at Washington

$$Q_{so}=39.9°+90.7 \cos (x+15.7°)-15.0 \cos (2x-13.4°);$$

with the snow continuously covering the ground during the period

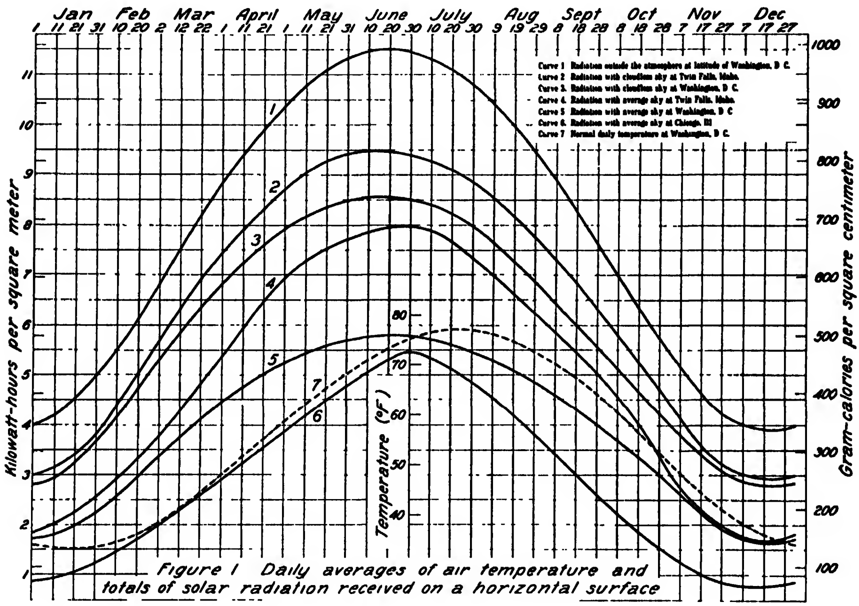

FIG. 22.—Daily averages of air temperature, and daily totals of solar radiation received on a horizontal surface.

December–February, and gradually disappearing during March, the series becomes

$$Q_s=10.9°+132.3 \cos (x-0.3°)-37.6 \cos (2x-49.4°).$$

The corresponding series for the temperature become

$$T_{so}=14.2°+10.5 \cos (x-15.5°),$$
$$T_s=10.0+15.3 \cos (x-15.5°).$$

Interpreting the above series, since the coefficient of reflection of snow is high, in the presence of a snow cover the *heat effective radiation* is greatly reduced during the winter. In consequence, winter temperatures are lowered about 9° C., or 16° F., below what they would be with no snow in the ground, and summer temperatures are not changed materially.

It is worthy of note that temperatures below zero have never been recorded at Washington when the ground was free from snow, while a minimum of $-15°$ F. has been recorded with the ground covered with snow.

The above results seem to indicate that changes taking place within the atmosphere are capable of producing greater temperature variations, and therefore greater weather changes, than can be brought about by solar variations of the order of magnitude indicated by researches that have been published up to the present time.

Solar radiation and atmospheric circulation.—It will be noted that for Washington the annual term for Q_T is plus. Ångström found that for Stockholm it was minus. If no heat were received except by radiation we would expect the annual term for Q_T to be zero. The latitude of Washington is 39° 54′ N., and of Stockholm, 59° 21′ N. Evidently, therefore, we find here evidence that a part of the radiant energy received in low latitudes is carried by convection to high latitudes. This is a clue to the energy available to keep the atmospheric engine running.[53, 54] Observations at a sufficient number of points ought to tell us how much energy is available for this purpose, and enable us to introduce this term into equations of atmospheric motion. This would be a long step in the direction of making meteorology an exact science.

At present we know that pressure distribution over land and sea gradually responds to changes in temperature distribution. Atmospheric motions result, which vary with the seasons, becoming most violent when the latitudinal differences are at a maximum. The motions would be comparatively steady were the Earth's surface frictionless and homogeneous in character throughout. Differences between the temperature over land and sea, uneven surfaces of land masses, and especially the rotation of the Earth, cause turbulent motions out of which result storms, cold waves, hot winds, and the whole train of variations in the meteorological elements which we call weather.

BIOLOGICAL EFFECTS OF SOLAR RADIATION

It is not within the province of meteorology to investigate biological problems. It will therefore suffice here to simply refer to some effects of solar radiation upon animal and vegetable life.

Primarily all life upon the earth is dependent upon the conversion of solar radiant energy received at the earth's surface into heat; for so far as we know some degree of heat is necessary to maintain life in any of its forms. But aside from this, neither plants nor animals, and especially those of the higher orders, will long exist if deprived of sunlight, even though their environment is maintained by artificial means at their

optimum temperature. However, the investigations of Spoehr [55] and others indicate that the various processes connected with plant development require such a small part of the available solar radiation that it is negligible as compared with the energy expended in the thermodynamic processes of the atmosphere. Also, Popp [56] has shown that plant development proceeds equally well when the ultra-violet light is excluded, as it is under ordinary glass in a green house. Most writers, however, maintain that radiation of all wave-lengths is necessary to the proper development and maintenance of plant life. Undoubtedly the light requirements of plants differ, depending upon the environment in which they were originally developed.

It seems to be well established that the higher orders of animal life will not thrive under glass, or in a smoky or hazy atmosphere, where the short-wave ultra-violet radiation is excluded. Hence, the efforts now being made to free the atmosphere of cities from smoke and other impurities, and to produce a window glass that transmits ultra-violet rays. Compare, for example, curve 6, Figure 22, giving the average daily totals of solar radiation at Chicago, with curves 4 and 5 for Twin Falls, and Washington, respectively. The deficit in Chicago is specially marked during the winter months when the conservation of solar radiant energy is most important.

THE DIRECT UTILIZATION OF SOLAR ENERGY

Referring again to Figure 22, from curve 2 we may compute that on a cloudless day in mid-summer at Twin Falls, Idaho, the daily receipt of solar radiant energy per square mile of horizontal surface is equal to nearly 33 million horse-power hours, and from curve 3, that at Washington it is equal to nearly 30 million horse-power hours. On an average day in midsummer the daily receipt at Twin Falls reaches 27.5 million, and at Washington about 20 million, horse-power hours.

This appears to be an enormous amount of energy, and comparable with the energy that may be derived per square mile of surface area from a storage lake in a water-power system. The difference lies in the fact that while the water-power may be developed at a single point, namely, the outlet of the lake, the solar energy is not so easily concentrated. The latter presents a problem analogous to that of utilizing water-power from many small ponds scattered over an extensive area.

The concentration of solar energy in a small way has been accomplished through the use of mirrors and utilized by Abbot in cooking.[57] He,[57] and also Ackermann,[58] refers to several devices that have been employed for the direct utilization of solar radiant energy, with varying degrees of success, but apparently without having attained economic efficiency.

In ages past solar radiation, by stimulating plant growth, has stored for our present use the supplies of coal and oil we are now spending so lavishly. When we recall that this storage was accomplished at a great cost in solar energy, ordinary caution should lead to careful conservation of what Nature has so bountifully provided, but which is by no means limitless in quantity. We may rest assured, however, that if, in the distant future, this supply of stored-up energy becomes exhausted, solar energy, the primary source, will still be unimpaired; and only man's ingenuity is required to make it directly and economically available in the form of power, heat or light, as required.

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CHAPTER IV

THE METEOROLOGY OF THE FREE ATMOSPHERE

Willis Ray Gregg, Leroy T. Samuels and Welby R. Stevens

INTRODUCTION

The "free atmosphere" is generally recognized as that portion that is out of range of surface recording instruments. The expression therefore ordinarily applies to all levels above about twenty meters, at or near which height anemometers and wind vanes are usually exposed. However, at some of the urban meteorological stations these and other instruments are placed on buildings which may be and in some cases are as high as 150 meters. Notwithstanding these occasional exceptions it seems proper to accept as surface data those within about twenty meters of the ground and as free air data those above that height.

The "meteorology of the free atmosphere" is an awkward expression and a single word "aerology" is in general use in its stead. It is usually defined as "the study of the free or upper air." Its derivative, "aerological," used in conjunction with "observations," "investigations," "data," etc., serves to define them as belonging to that portion of the atmosphere immediately above the surface layer.

Several methods are in general use for making observations of the upper air each of which has certain advantages depending on the weather conditions encountered, the particular elements to be observed, the heights to be reached, etc.

For some interesting historical facts regarding aerological observations the reader is referred to Chapter II, pages 30-31, of the section on the meteorology of the free air.

METHODS OF UPPER AIR OBSERVATIONS

Clouds.—(a) *Eye observation.*—Since clouds occur at all levels from the surface of the earth up to about 11 km. above sea-level in the middle latitudes and occasionally as high as 15 km. in the tropics, eye observations of their movements provide the simplest means of determining the wind direction at those elevations. If the height of the clouds is known, then a fairly accurate determination of the velocity also can be made by an experienced observer. For a description of typical cloud forms, with illustrations, see Cloud Classification, Chapter II, and Figures 1-6.

(b) *Nephoscope.** This instrument consists chiefly of a black mirror, mounted in a circular frame which is graduated in degrees, and a movable sighting device containing two eyepieces at selected heights, viz., $166\frac{2}{3}$ mm. and $83\frac{1}{3}$ mm. above the surface of the mirror.[1] The nephoscope is mounted on a level stand which is set out in the open where the mirror will reflect as much of the sky as possible (Figure 23).

Cloud directions can be obtained with a high degree of accuracy by this means, and when their height is known the wind velocity at that level can likewise be accurately determined. The motion of the cloud image as

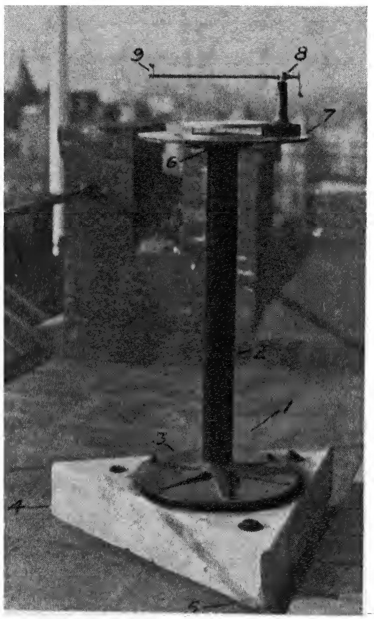

FIG. 23.—Nephoscope, complete with support.

seen in the mirror indicates the direction in which the clouds are moving, and from the distance the image moves in a given time and from the cloud height the velocity is readily computed.

Pilot balloons.—To determine the wind direction and velocity at various elevations at times of no clouds, or at levels below the clouds, use is made of pilot balloons. This method consists mainly of inflating a 6″ rubber balloon with hydrogen gas to approximately 24″ diameter, releasing and then following it by means of a theodolite (modified surveyor's transit)

* For a more detailed description of this and of other instruments here described, the reader will please consult the references given.

from which the elevation and azimuth angles are read each minute. The underlying basis of this method is the assumption that the balloon rises at a uniform rate when inflated to a definite "lift." The ascensional rates for various "lifts" were determined from a large number of observations made with two theodolites.[2] The proper "lift" in single theodolite observations is secured by means of a specially adapted weighing balance after taking into account the weight of the balloon[3] (Figure 24). From

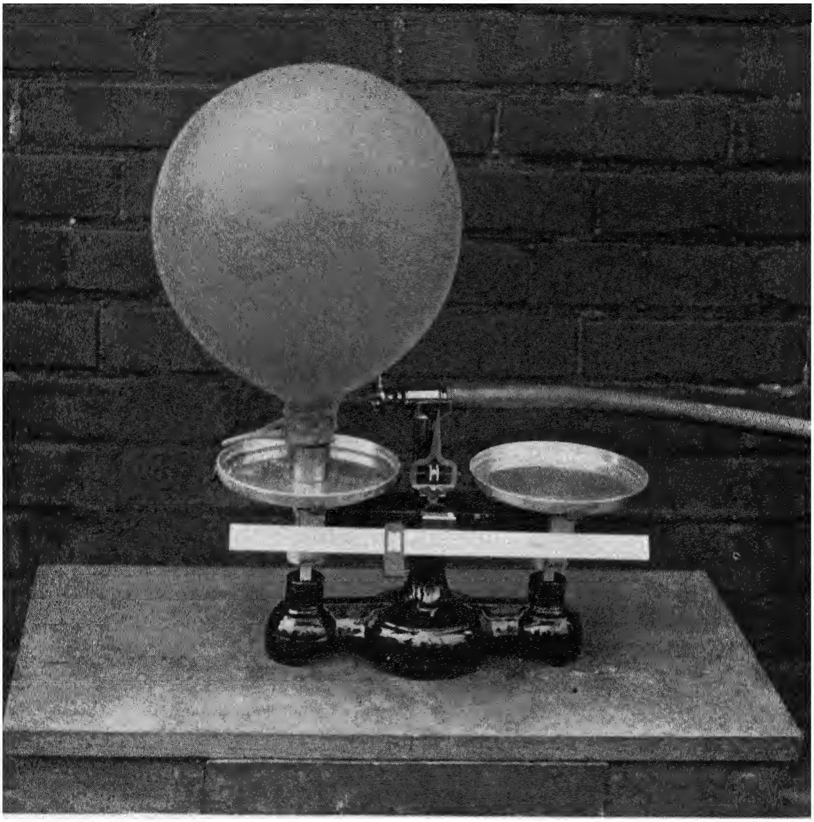

FIG. 24.—Inflation balance used for definite "lift."

the angular readings thus obtained the wind direction and velocity at the height of the balloon each minute are readily computed.

At most Weather Bureau stations these computations are completed between each two minutes' readings. For this purpose a telephone system is used whereby the observer at the theodolite reads and telephones the angular values to a computer who performs the necessary operations with the aid of a slide-rule and specially constructed protractor[4] (Figures 25A and 25B).

Balloons of certain colors are used so as to secure the best contrast against the sky background. In general, uncolored (pure gum) balloons

FIG. 25A.—Pilot balloon observation, showing observer following balloon with theodolite.

are best on days with no clouds, red balloons with high clouds and black balloons when clouds are low and dark.

Pilot balloon observations are also made at night by attaching to the balloon a small paper lantern containing a lighted candle or a small

electric light. The light is then followed instead of the balloon, but naturally this cannot be observed to as great heights as when the balloon is followed in the daytime. Night observations generally are limited to about 3 km. whereas in the daytime the balloon can be followed to heights of 10 to 15 km. or more, depending on the visibility and on the horizontal distance the balloon is carried by the wind. Occasionally these distances

FIG. 25B.—Computer at plotting board inside the office in telephonic communication with observer.

are very great. For example, on November 16, 1925, at Broken Arrow, Oklahoma, with two theodolites, a balloon was followed until its horizontal distance was 63,750 meters, its height being 6,900 meters. At this point the balloon was lost by one of the theodolites but, by assuming a continuation of the same ascensional rate, it was followed with the remaining theodolite until its horizontal distance was 87,250 meters and its height 8,200 meters. On July 3, 1926, at this same station, with two theodolites, a pilot balloon was followed to a height of 21 km. A uniform ascensional

rate of 190 meters per minute was maintained up to 19 km., but between 19 and 21 km. it decreased abruptly to about 100 meters per minute.

Kites.—Where upper air temperatures, pressures and humidities are desired, box-kites are used under certain wind and weather conditions. This method, however, is practicable only in winds between 5 and 35 meters per second, the maximum limiting velocity varying inversely as the air density and therefore increasing with height above sea-level.

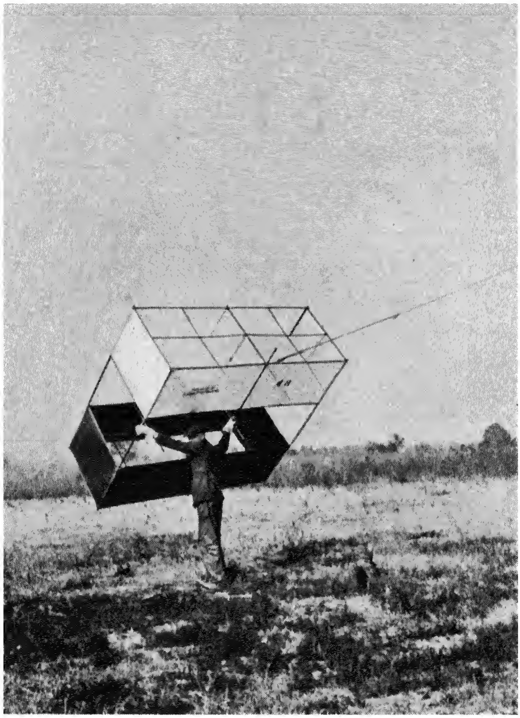

FIG. 26.—Launching a kite.

The Weather Bureau uses Marvin-Hargrave kites, which are made in three principal sizes approximating 7′ x 7′ x 3′.[5] (Figure 26). These kites are flown in tandem by means of steel piano wire. Usually four or five but occasionally seven or eight kites are used in a flight, depending on the amount of pull exerted on the wire, the latter being constantly indicated by a dynamometer mounted on the kite reel.

The Marvin meteorograph is used by the Weather Bureau in its kite work.[6] This instrument records temperature, pressure, humidity and wind velocity (Figures 27A and 27B). The wind direction aloft is obtained by eye observation on the kites. Average heights of nearly 3 km. are attained, and individual flights occasionally exceed 7 km.

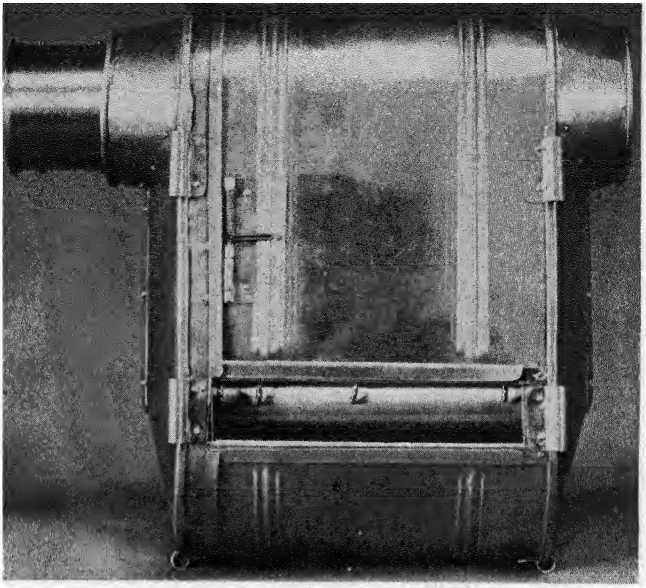

FIG. 27A.—Marvin kite meteorograph, complete, with cover.

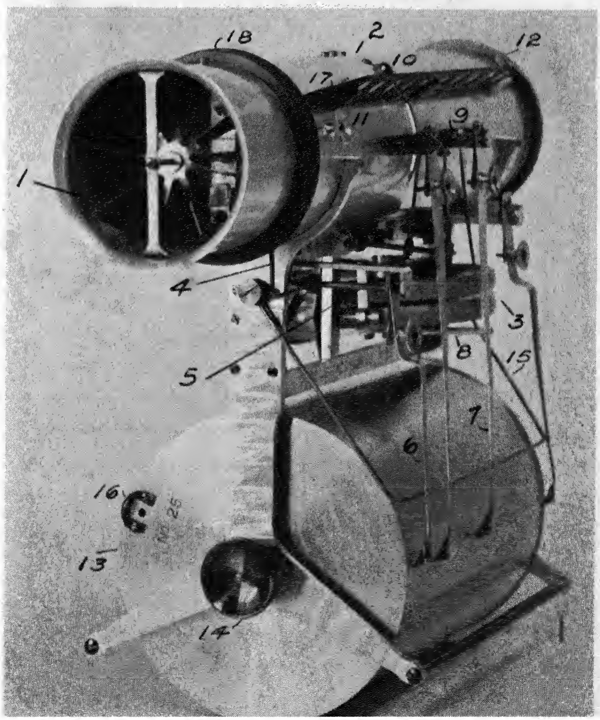

FIG. 27B.—Marvin kite meteorograph, complete, with cover removed.

The kite reel [9a] is made largely of cast iron and is of sufficient weight to render it stable under any pull exerted by the kites (Figure 28). The tensile strength of the kite wire is from 350 to nearly 500 lbs., depending on the diameters which are, 0.036″, 0.040″, and 0.045″, respectively. The largest size wire is placed on the reel first and therefore will be nearest the ground where the pull is greatest. Occasionally breakaways occur, owing to defects in the wire or to excessive pulls exerted by strong winds encountered aloft. The reel is mounted in a specially constructed house which can be turned to face in any direction (Figure 29). Suitable

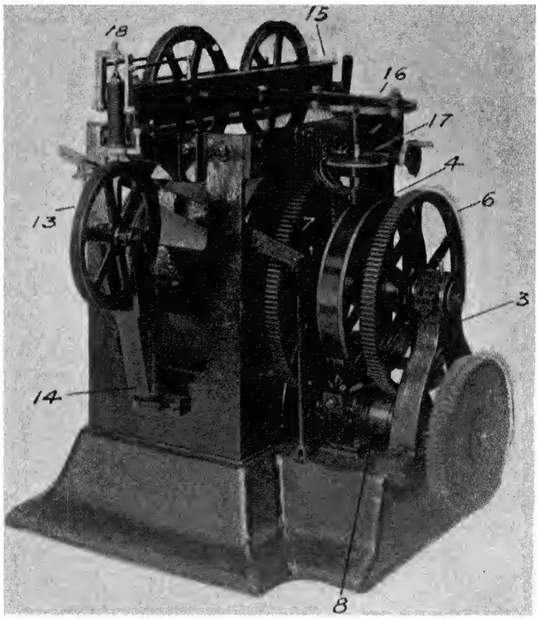

FIG. 28.—Kite reel.

ground connections are installed to provide protection against lightning and to carry off the electric potential which at times becomes dangerously high. The reel is operated by a variable speed, 4 h. p., electric motor.

Airplanes.—This method of upper air observation has become highly practicable and possesses many important advantages over kites. Chief of these is the short time required for a flight. An airplane can carry a meteorograph to a height of 3 to 4 km. and return to the ground in a half hour or less, whereas the average kite flight to the same altitude requires about four hours. Another important advantage is that an airplane can rise in a calm or very light wind, whereas a kite requires a wind

of at least 5 meters per second. However, kites have the advantage of being able to fly in a fog, whereas it is dangerous, as yet, for an airplane to take off in a fog unless there is assurance of a landing place which is free from fog.

Heights of 6 km. or more are practicable with airplanes and the records obtained are very satisfactory. The aero-meteorograph (Friez type used in this country) is mounted in a specially constructed frame on the wing of the airplane and records temperature, pressure and humidity.

Sounding balloons.—For heights greater than those possible with kites or airplanes, use is made of sounding balloons. These may be of two types, viz., those which are left open so that the gas can escape as the balloon rises and those which are sealed and expand as they rise

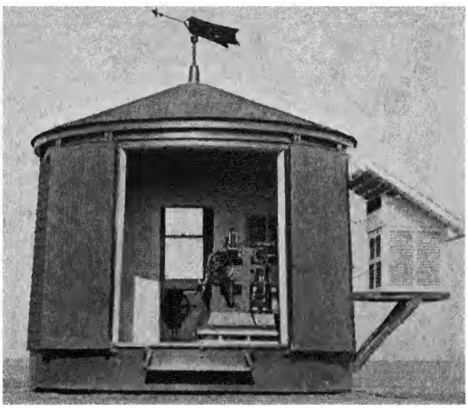

FIG. 29.—View of reel house, showing interior.

and finally burst. The latter are now used by the Weather Bureau (Figure 30).

Sounding balloons may range in size from less than one meter to more than ten meters in diameter. These balloons carry a meteorograph and usually attain heights between 15 and 20 km. and occasionally above 30 km. In this connection L. H. G. Dines has recently published an article entitled, "The Maximum Heights Which It Is Possible to Reach by Sounding or Pilot Balloons."[7] He concludes that elevations above 30 km. are practically prohibitive in view of the relative expansive qualities and corresponding weight of the rubber.

The instrument's descent is retarded by means of a small parachute. An attached card requests the finder to pack and return the instrument by express (collect), for which service a payment of from one dollar to five dollars is usually made.

Fig. 30.—Sounding balloon and meteorograph attached.

The percentage of returned instruments in this country is usually between 80 and 90. The majority of the instruments land within 50 miles of the station although many land at less than half that distance.

The Fergusson balloon meteorograph is used for sounding balloon work in this country [8] (Figure 31). In addition to the attainment of greater heights with sounding balloons than with kites there is the additional advantage that the former can be used in both lighter and stronger winds as well as during the occurrence of heavy precipitation. The chief dis-

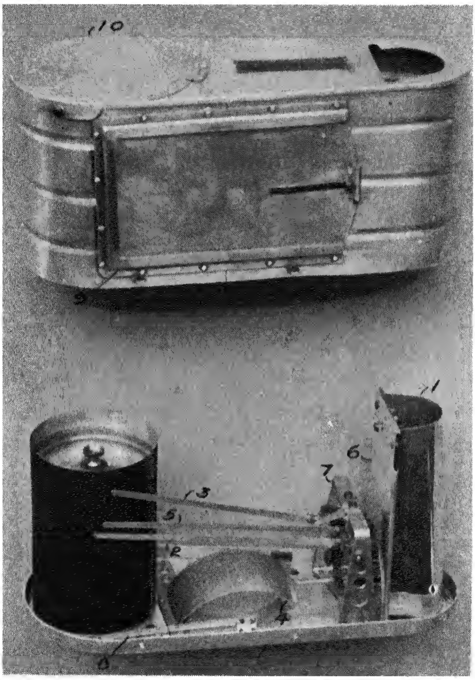

FIG. 31.—Fergusson balloon meteorograph, with cover above.

advantage of sounding balloons is that the records are frequently not available until days, weeks or perhaps months later.

Limited-height sounding balloons.—In this method the regular sounding balloon equipment is used, but at some predetermined height the gas is allowed to escape from the balloon, or the meteorograph is automatically detached and descends while the balloon escapes. The former scheme is employed by the Weather Bureau and is accomplished by means of a Rossby deflation valve. The valve is held closed by means of a thread and spring. The thread is placed through one end of a piece of fuse

which burns at a known rate. The other end of the fuse is lighted when the balloon is released. When the fire reaches the thread the valve opens and the gas escapes (Figure 32).

The second method is used mostly in Europe and can be accomplished with a special device known as the "Baker automatic release" which is actuated by an aneroid cell.[9] The first method, however, has the advantage of permitting the recovery of the balloon.

Limited-height sounding balloons are especially suitable for sparsely settled regions or over water areas where the chances of having the instrument found and returned are slight. When used over water a second balloon and a floating device are attached. This extra balloon serves to keep the instrument above the water for a few hours after its descent.

Radio sounding balloons.—Capt. Robert Bureau of the French meteorological office has recently performed experiments by suspending from a sounding balloon a light, short-wave radio-telegraphic transmitter whereby signals may be received at the ground during the ascent of the balloon even after the latter has entered the stratosphere.[10] The apparatus is designed to give indications of the pressure and temperature elements.

The first sounding for temperature was made on January 17, 1929, and several others have been made since. A light baro-thermograph was attached to the outfit to verify the transmissions of the radio-thermometer. With a temperature range from 20° to −60° C. the error was found to be less than 0.7°.

This method allows almost immediate knowledge of the free-air data and therefore has decided advantages in this respect.

Radio pilot balloons.—Experiments are being made by the Signal Corps, U. S. Army, in the use of pilot balloons during the prevalence of fog and low clouds. The method which has been developed depends primarily on a small radio transmitter, weighing less than one pound, attached to a cluster of three ordinary pilot balloons. The latter are assumed to rise at a constant rate, and from radio signals transmitted the wind direction and velocity at various elevations are determined.

The signals are picked up by means of a specially designed receiving set with a loop antenna, similar to that used in the radio compass or direction finder.

In the course of an observation two, or preferably three, of these radio-direction finders are set up at the ends of baselines of known length, which are arranged to form a triangle when three sets are used. From regular one-minute interval reading of the signals transmitted, the balloon's azimuthal bearing is readily determined.

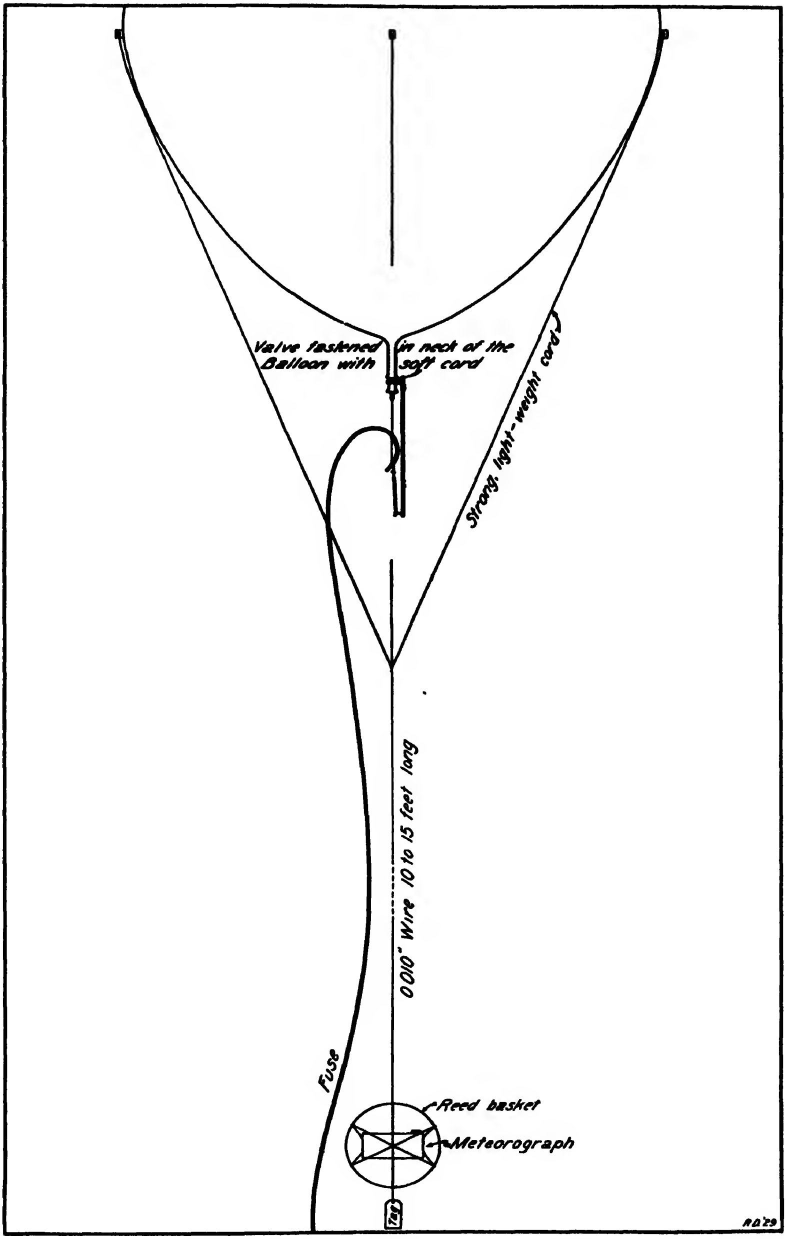

FIG. 32.—Limited-height sounding balloon equipment showing Rossby deflation valve attached to neck of balloon.

Captive balloons.—This method is especially suitable when winds are too light for kite flying. Generally one or more large balloons (20 to 40 cu. meters) are used. These are fastened to a wire or cable which is allowed to unwind as the balloon rises. The greatest heights are attained during the lightest winds, which do not carry the balloon far away horizontally and the amount of wire required to be lifted is less. The meteorograph is hung about 5 meters below the balloon. Heights of

FIG. 33.—Free-rising captive-balloon.

3 to 4 km. can be obtained under favorable conditions and, by using two or more balloons, heights above 6 km. are possible.[11]

Free-rising captive balloons.—In this method instead of the balloon pulling its line out as does a captive balloon and kite, the wire is reeled out so rapidly that the balloon rises free from all restraint except that of the gradually increasing weight of the wire. The reel is mounted on a platform about 10 feet high, so that the slack wire will not come in contact with the ground. Heights of 3 km. or more are attainable by this method (Figure 33).

Kite balloons.—These are used chiefly by military services for observation purposes. They are usually intended to carry one or more men but are sometimes used for carrying only meteorological instruments. They are cylindrical with hemispherical ends and are attached to a cable. The cylinder is divided by a diaphragm near its lower end into two chambers, the upper and larger of which is filled with gas, while the lower one, by means of a valve opening inwards, receives the pressure of the wind which presses against the diaphragm and preserves the sausage form of the balloon in spite of the leakage of gas. Another wind-bag encircling the bottom of the air chamber serves as a rudder, and lateral fins or wings give stability to the balloon about its longer axis. The meteorograph is hung at a considerable distance below the balloon.[12]

Ceiling balloons.—These constitute a recent development and are used mainly in connection with commercial aeronautics to determine cloud heights when the latter are relatively low. Small rubber balloons (4 x 6 in.) inflated to about 15 x 20 in. in diameter are given a definite "free lift" (40 grams) and rise at a practically uniform rate of about 100 meters per minute. They are followed by eye and timed until they enter the cloud base. This method provides an inexpensive and accurate means of determining cloud heights when the latter are low, *i. e.*, less than 1,000 meters.

Ceiling lights.—This method is used to determine cloud heights at night and consists of projecting a spot of light onto the cloud base. In some cases the light is projected vertically and in others it is set at some definite angle, *e. g.*, 45°, 63° 26′, etc. The height of the light spot on the cloud is determined by means of an alidade, or protractor. The angular reading of this instrument together with the known distance of the light source from the point of observation provides the necessary data. The alidade is usually graduated in heights instead of degrees, thus eliminating the necessity of computation or the use of tables or graphs. This method is now in use primarily for aviation purposes.

Free-manned balloons.—This method provides a means of determining the path, or trajectory, of air moving in response to given pressure gradients. Such a series of flights was carried out by the late Dr. C. L. Meisinger.[13] Ten flights were made from Scott Field, Illinois, between April 1, and June 2, 1924. The records obtained during this series of flights have been summarized by V. E. Jakl.[14]

Archie bursts.—This method of determining the wind direction and velocity of upper winds consists of sending up a bomb or shell which bursts at a predetermined height. The drift of the smoke cloud from the burst is watched in a specially designed mirror or pair of mirrors. When two mirrors are used the height of the smoke puffs can be determined,

whereas with a single mirror the height must be assumed in order to compute the wind velocity.[15]

Another method in which shell bursts are employed is used for the determination of wind direction and velocity above a cloud sheet. A number of shells (4 to 6) are caused to burst at known intervals of time (say every 60 secs.) at the heights required. The smoke from the successive bursts is carried along by the wind, the average distance between each puff being $60v$, where v is the average wind velocity. If then, an airplane whose speed is V, flies along the line of smoke drift and the average time taken to pass successive puffs is, t,

$$Vt = 60v,$$

or

$$\therefore v = \frac{Vt}{60}.$$

The wind direction is deduced from the direction of flight of the airplane as read from the compass.[15]

Sound ranging.—This method consists of sending up small balloons loaded with bombs which burst after a certain time, the position of the bursting being determined by sound ranging from the ground. The average rate and direction of travel of the balloon up to the level of the burst is thus obtained. By arranging for several bombs to burst at different heights the average wind movement in each intermediate layer can be determined.[15]

Noctilucent clouds.—This phenomenon consists of luminous night clouds at altitudes considerably above those of ordinary clouds. They are characterized by their brilliant prismatic colors and are attributed to the sun's illumination of a stratum of foreign matter which may have been carried aloft by volcanic eruptions. Their greatest brightness occurs not far above the horizon in a clear sky in the twilight segment. Photographic observations of these clouds, particularly if from two stations simultaneously, are of great importance, because they provide a means of determining wind movements at heights which could otherwise be obtained only from meteor trails and auroras.[16]

The most recent determinations of the heights of noctilucent clouds were made by Prof. C. Störmer when on January 13, 1929, he secured more than 90 simultaneous photographs from two stations in Norway.[17] Some of the photographs were taken late at night when the stars were visible. The heights could then be determined with considerable accuracy, since the distance between the two stations was 27 km. The average altitude of the clouds was found to be about 25 km., which was in close agreement with the values he found from two pairs of similar observations that he made on December 30, 1926.

According to O. Jesse,[16] who observed these clouds a number of times between 1885 and 1894 between latitudes 45° and 64° N., the illuminated portions occur at an altitude of about 82 km. His measurements gave velocities of between 25 and 200 meters per second. It is interesting to note that in none of his observations did the clouds move from a direction between southeast and southwest. In most cases the movement was from the northeast and east-northeast, with a secondary maximum from west-southwest.

Archenhold [16] also calculated a height of 80 km. from observations made by him in 1910, although based on photographs from only one station.

Meteors.—Meteors afford direct evidence of the existence of the earth's atmosphere at the heights at which they are observed, since the glow we see is caused by the swiftly moving body (presumably of cosmic origin) becoming heated to incandescence by friction with the atmosphere. They are most accurately observed by means of photographs, and when secured from two stations simultaneously it is often possible to determine not only the height of the path but also the linear velocity.[18]

According to Denning the average heights at which meteors appear and disappear is 122 km. and 82 km. respectively.[19] The most frequent limits of their appearance are between 160 and 90 km. and they usually disappear below 120 km. (mostly at 80 km.) Their maximum height of appearance is about 200 km. Their velocities average about 43 km. per second, and range from 10 to 160 km. per second.

Auroras.—Observations of auroras furnish evidence of the vertical extent of the earth's atmosphere to greater heights than does any other measurement. Simultaneous photographs of the same aurora from two stations against the same background of stars provide a practical means of determining the height with considerable accuracy. According to Störmer,[20] and Vegard and Krogness,[21] who have made a large number of measurements of these heights, it is found that the lower edge of the aurora, which is usually sharply defined, averages between 100 and 110 km. in height. Heights exceeding 200 km. are occasionally observed and on March 22-23, 1920, Störmer measured several heights exceeding 600 km. and one of 750 km. Recently he has found that diffuse auroras extend to heights of more than 1,000 km.[22] and also that auroras of great height (over 400 km.) seem to occur only in sunshine.[23]

At their upper limits auroras often fade away gradually and the height to which they can be traced in a photograph depends largely upon the sensitiveness of the plate.

Goddard rockets.—This method is still in the experimental stage. It was developed by R. H. Goddard, and is based on the assumption that

if nitrocellulose smokeless powder were employed as propellant in a rocket, under certain conditions an efficiency of 50 per cent might be expected.[24] Determinations were made of the mass which would be required to attain heights of 56, 116 and 373 km., respectively, as well as to reach sufficiently high altitudes to escape the earth's attraction.

On account of the great speed with which this rocket would rise (57km./min.) and likewise the great speed with which it would fall, at least until it reached air of sufficient density to retard it appreciably, special apparatus would need to be developed for recording meteorological conditions. A type of such a device has been suggested by S. P. Fergusson which contains a pressure and temperature element.[8]

VERTICAL STRUCTURE OF THE ATMOSPHERE

The physical changes which occur in the free atmosphere are discussed from a theoretical viewpoint in Chapter V which follows. In this present section are given (mostly graphically) various average values as determined from observations made principally in the United States.

Annual march of free-air temperature.—Free-air data based on kite observations made during the past 11 years at Ellendale, North Dakota, Broken Arrow, Oklahoma, Groesbeck, Texas, and Royal Center, Indiana, 10 years at Drexel, Nebraska, and 9 years at Due West, South Carolina, suffice to give average values at these stations. Since Ellendale, Drexel, Broken Arrow and Groesbeck all lie close to the same meridian (97th), and nearly equally spaced between the 31st and 46th parallel, their data furnish information regarding the latitudinal variation of the various elements. The other two stations, viz., Royal Center and Due West, provide representative data for the region just south of the Great Lakes and in the south Atlantic Coast States, respectively.

In considering the following averages it should be understood that the results are based on kite records and, therefore, are representative of conditions when kites can be flown, *i. e.*, during winds of between 5 and 35 m. p. s. and when precipitation is not occurring at an appreciable rate. It happens, however, that the average times of kite flights, viz., from shortly after sunrise to about noon, is such that the mean values of the meteorological elements at the surface are very nearly the same as the 24-hour averages. The differences are, in general, very small, and have previously been published in detail.[25]

Figures 34, 35, and 36, for Ellendale, Drexel and Groesbeck, have been selected to show variations in the annual march of free-air temperatures with latitude and season of the year, and Figure 37 for Due West to show differences between free-air temperatures over the plains of Texas

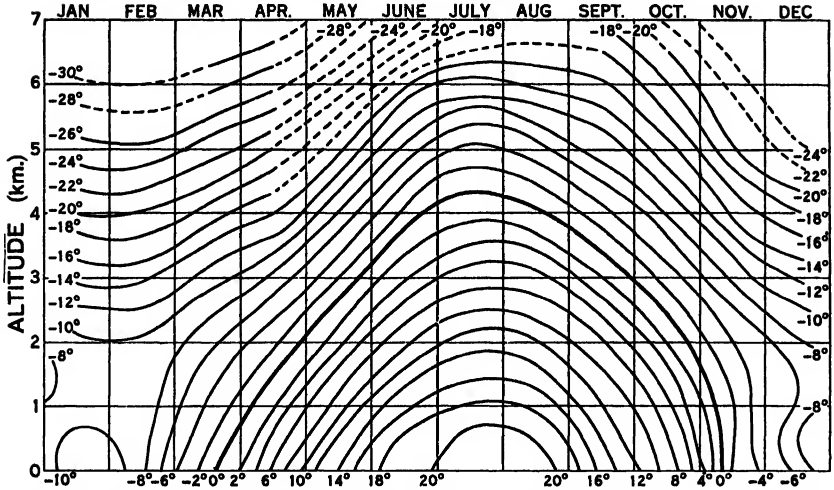

Fig. 34.—Mean annual march of free-air temperatures, °C., above Ellendale, N. Dak., lat. 45° 59′ N., long. 98° 34′ W., elevation 444 meters. (Lines extended to sea-level by extrapolation).

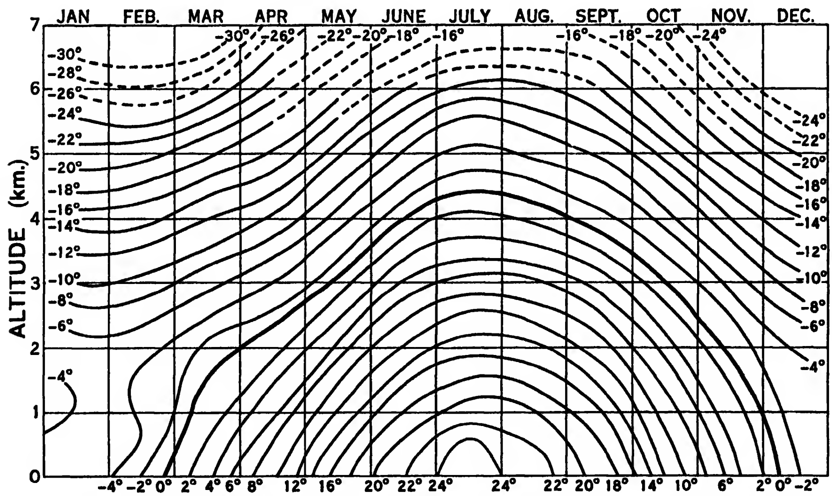

Fig. 35.—Mean annual march of free-air temperatures, °C., above Drexel, Nebr., lat. 41° 20′ N., long. 96° 16′ W., elevation 396 meters. (Lines extended to sea-level by extrapolation).

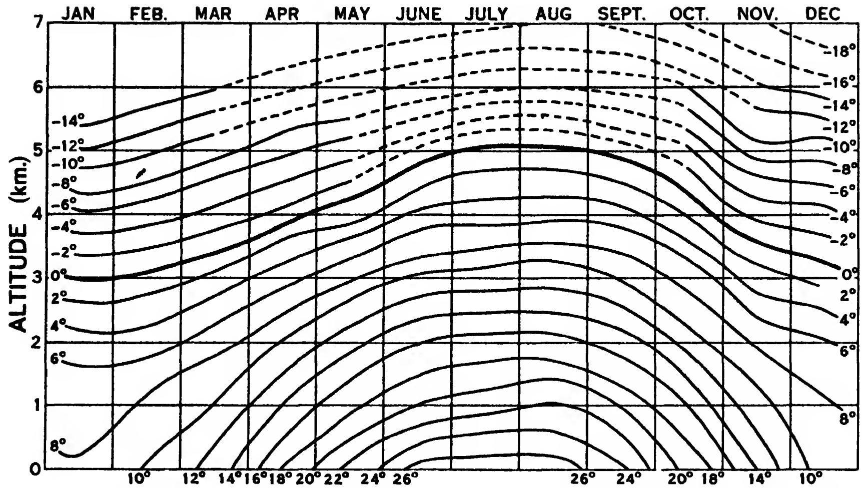

FIG. 36.—Mean annual march of free-air temperatures, °C., above Groesbeck, Tex., lat. 31° 30′ N., long. 96° 28′ W., elevation 141 meters. (Lines extended to sea-level by extrapolation).

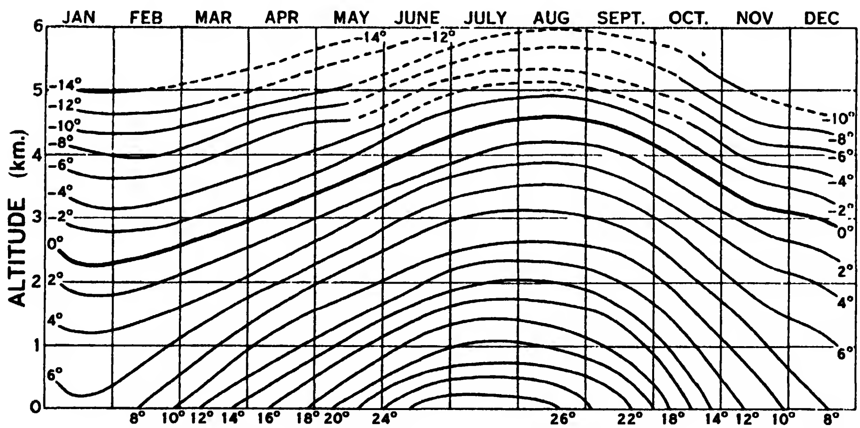

FIG. 37.—Mean annual march of free-air temperatures, °C., above Due West, S. C., lat. 34° 21′ N., long. 82° 22′ W., elevation 217 meters. (Lines extended to sea-level by extrapolation).

and near the Atlantic Coast. Note particularly the inversion of temperature in winter at Ellendale, the northernmost station, from the surface to 2 km. elevation, the almost uniform temperature between these altitudes at Drexel, and the low corresponding lapse rate at Groesbeck as compared with that at Due West. Also, the annual amplitude in the monthly means of temperature decreases from north to south at all altitudes. It is less at high than at low levels. Thus, at Groesbeck, at the surface the annual amplitude is 59 per cent of that at Ellendale, and at the 5-km. level, 54 per cent. Due West, likewise, has a comparatively small amplitude.

Since sufficient sounding balloon observations have not been obtained in the United States to make possible the construction of graphs representing conditions at great heights, observations made in the British

TABLE 6

HIGHEST AND LOWEST MONTHLY MEAN TEMPERATURES (° C.) AT THE SURFACE AND AT 5 KM. ABOVE SEA-LEVEL

Station	Surface			5 km.			Percentage of amplitude at 5 km. of that at surface
	Maximum	Minimum	Amplitude	Maximum	Minimum	Amplitude	
Ellendale	21.0 Jul.	—11.1 Jan.	32.1	— 3.6 Jul.	—25.6 Feb.	22.0	68
Drexel	24.6 Jul.	— 6.0 Jan.	30.6	— 3.6 Jul.	—24.8 Feb.	21.2	69
Broken Arrow	26.7 Aug.	3.2 Jan.	23.5	— 1.1 Jul.	—15.3 Feb.	14.2	60
Groesbeck	26.8 Aug.	8.0 Jan.	18.8	0.3 Jul.	—11.5 Jan.	11.8	63
Royal Center	24.9 Jul.	— 4.3 Jan.	29.2	— 3.6 Aug.	—23.2 Feb.	19.6	67
Due West	26.8 Jul.	5.7 Jan.	21.1	— 0.5 Aug.	—14.5 Feb.	14.0	66
England	16.0 Aug.	3.0 Jan.	13.0	—11.0 Aug.	—24.0 Feb.	13.0	100

Isles [26] have been utilized to construct Figure 38. A number of characteristics of a typical marine climate are conspicuous on this chart. Thus, the annual amplitude of the mean monthly temperature, both at the surface and aloft, is considerably less than over American stations. The mean winter temperature at the surface for England (Latitude 51° N.) is practically the same as for Broken Arrow (Latitude 36° N.), while at 6 km. elevation the temperature is lower than over Ellendale (Latitude 46° N.) at this altitude, conforming to the true latitudinal relationship between the two regions.

In Table 6 are given the highest and lowest monthly mean temperatures for the surface and for 5 km. above sea-level. The last column of this table is of special interest in that it shows some striking relationships between the amplitude of the average annual temperature at the surface and that at 5 km. It will be noted that the amplitude at 5 km. is practically two-thirds that at the surface at all the American stations,

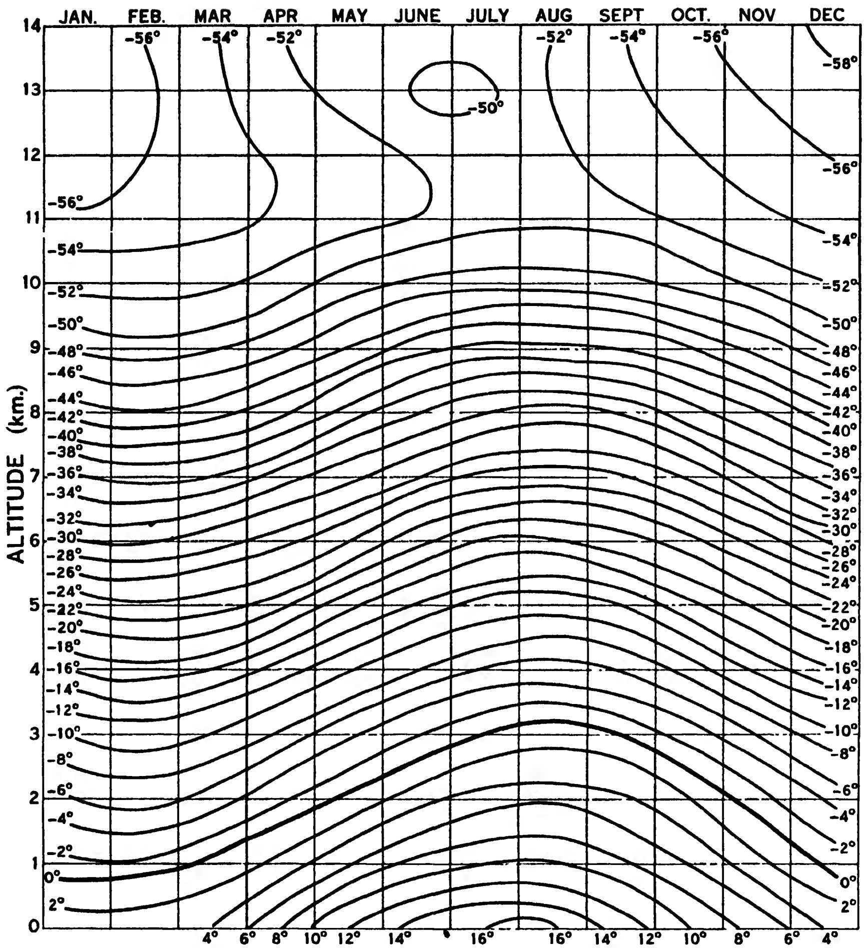

FIG. 38.—Mean annual march of free-air temperatures (°C.) above the British Isles.

and is slightly greater at the northern than at the southern stations. Over England, however, the annual range at 5 km. is the *same* as that at the surface. Above that altitude, the range gradually diminishes until, at and above 10 km., it is approximately one-half that at the surface.

From this table it is also seen that the annual maximum temperature occurs during the same month (July) at the 5-km. level and at the sur-

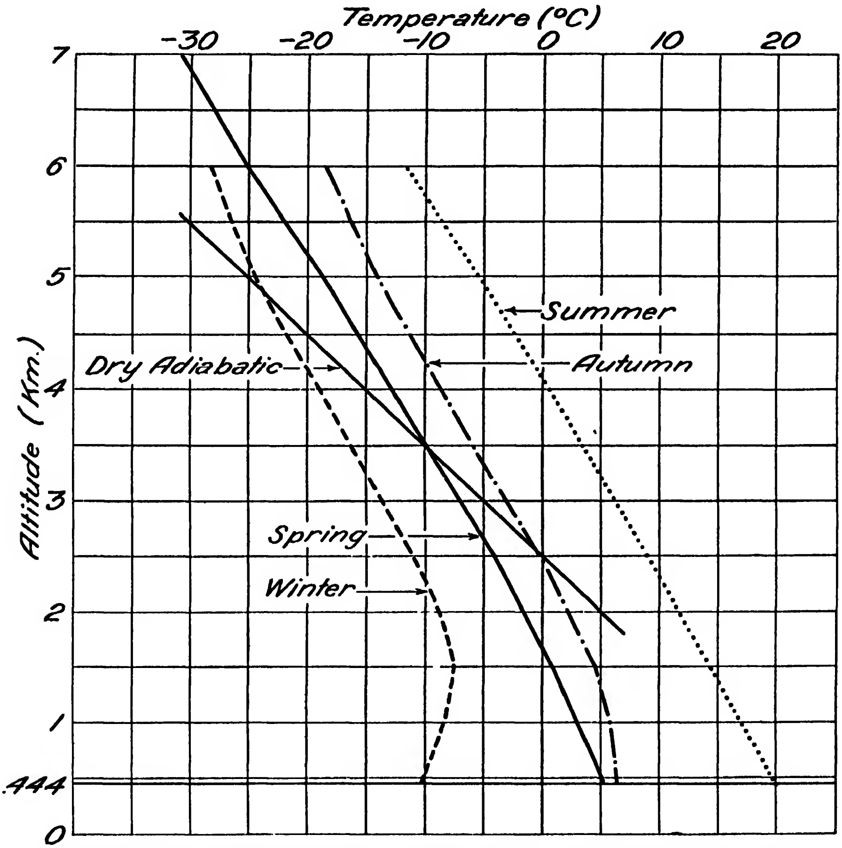

FIG. 39.—Mean seasonal free-air temperatures (°C.) at Ellendale, N. Dak.

face over Ellendale, Drexel, and England; it occurs one month later aloft than at the surface over Royal Center and Due West, and one month earlier aloft than at the surface over Broken Arrow and Groesbeck. The minimum occurs at the surface and at the 5-km. level during the same month (January) at Groesbeck and one month later at the upper level than at the surface at all other places.

Mean free-air temperatures.—Figures 39, 40, and 41 show the mean seasonal free-air temperatures (° C.) for Ellendale, Groesbeck and Due

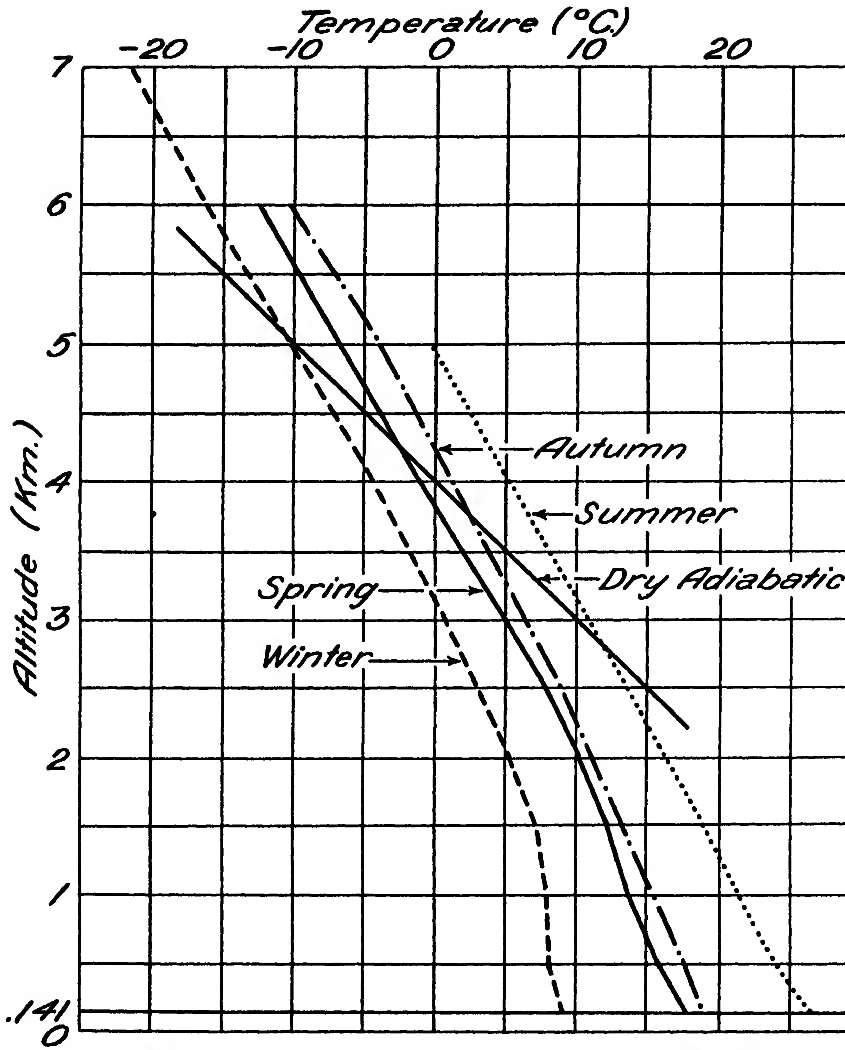

FIG. 40.—Mean seasonal free-air temperatures (°C.) at Groesbeck, Tex.

West for the same period as represented by Figures 34, 36, and 37. These are, respectively, the most northern, southern and eastern kite stations in the United States.

Attention is invited to the following:

1) The decided permanent inversion of temperature in the lower levels at the northern stations during the winter.

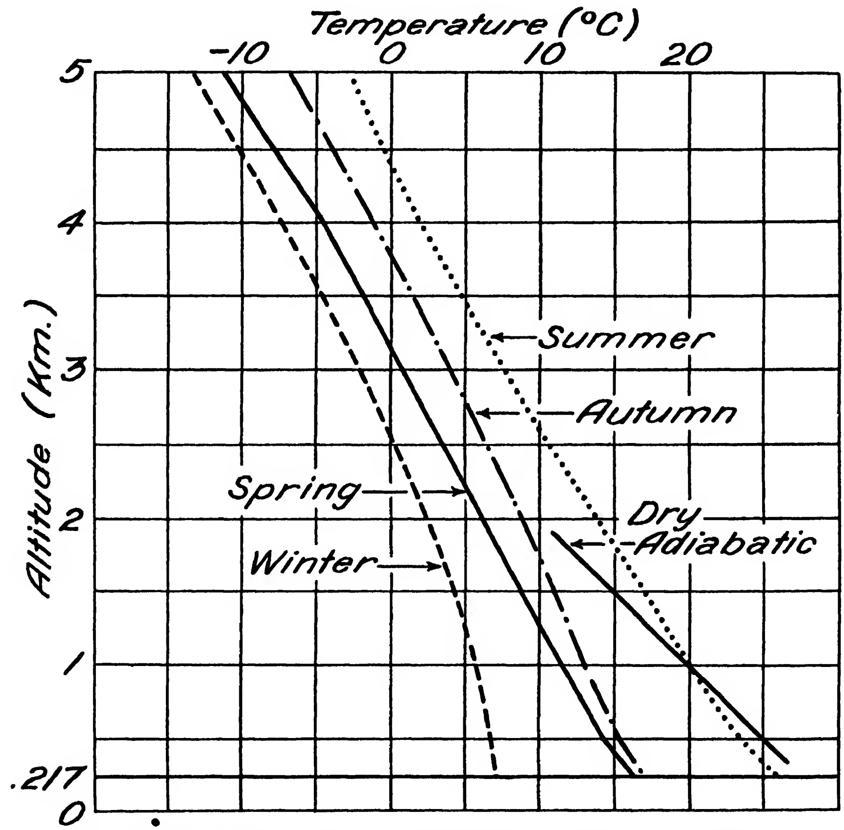

FIG. 41.—Mean seasonal free-air temperatures (°C.) at Due West, S. C.

2) The large latitudinal difference in annual temperature range, both surface and aloft.

3) The divergence of the spring and autumn curves with increase in altitude, the former approaching the curve for winter and the latter approaching that for summer. This is due to the fact that the air nearest the ground becomes warmed more rapidly than at higher levels with the advance of spring, and conversely cooled in the same way with the advance of autumn.

In Table 7 are given the mean lapse rates for summer and winter at the three above-named stations.

The data are based on the same periods of records as those shown in Figures 34, 36, and 37.

The following features are shown:

1) The mean lapse rates are greater in summer than in winter.

2) The mean lapse rates in the lower layers are less at the northern than at the southern and eastern stations.

3) The mean lapse rates in summer are greater in the lower layers than in the upper and vice versa in winter.

TABLE 7

MEAN LAPSE RATES FOR SUMMER AND WINTER AT ELLENDALE, NORTH DAKOTA (444 M.), GROESBECK, TEXAS (141 M.), AND DUE WEST, SOUTH CAROLINA (217 M.)

Altitude (meters) M. S. L.	Summer			Winter		
	Ellendale	Groesbeck	Due West	Ellendale	Groesbeck	Due West
Surface						
500	0.53	0 75	0.98	+0.18	0.22	0.18
1,000	0.50	0.46	0.64	+0.38	0.04	0.20
1,500	0.50	0.50	0.64	+0.12	0.20	0.30
2,000	0 56	0.52	0.64	0.26	0.34	0.36
2,500	0.58	0.56	0.62	0.44	0.44	0.42
3,000	0.58	0.56	0.58	0.52	0.46	0.46
4,000	0.57	0.55	0 58	0.54	0.52	0.53
5,000	0.55	0.55	0.44	0.57	0.57	0.60

4) The mean lapse rates are in close agreement above the 2,000 m. level at all stations.

5) The mean lapse rates are inverted (positive) in winter at Ellendale from the surface to the 1,000 m. level.

Mean free-air temperature at great heights.—In Figure 42 are shown the mean free-air temperatures at places of various latitudes based upon the data published by Shaw.[27] The curve for the United States is taken from Gregg.[28]

Attention is invited to the following features:

1) The decrease in the height of the base of the stratosphere (tropopause) with increase in latitude.

2) The tropopause is at a higher elevation in summer than in winter.

3) The variation in the mean temperature between widely separated places persists to about 10 km. At 12 km. the mean temperatures for all of the stations closely approach the same value.

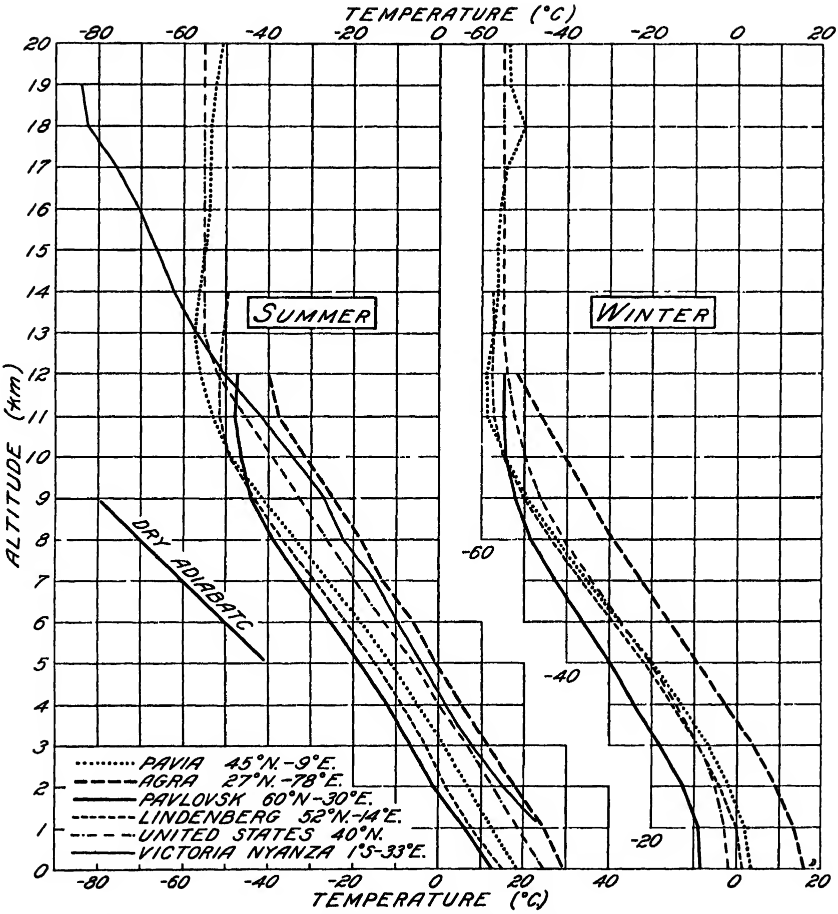

FIG. 42.—Mean free-air temperatures for places of various latitudes for summer and winter.

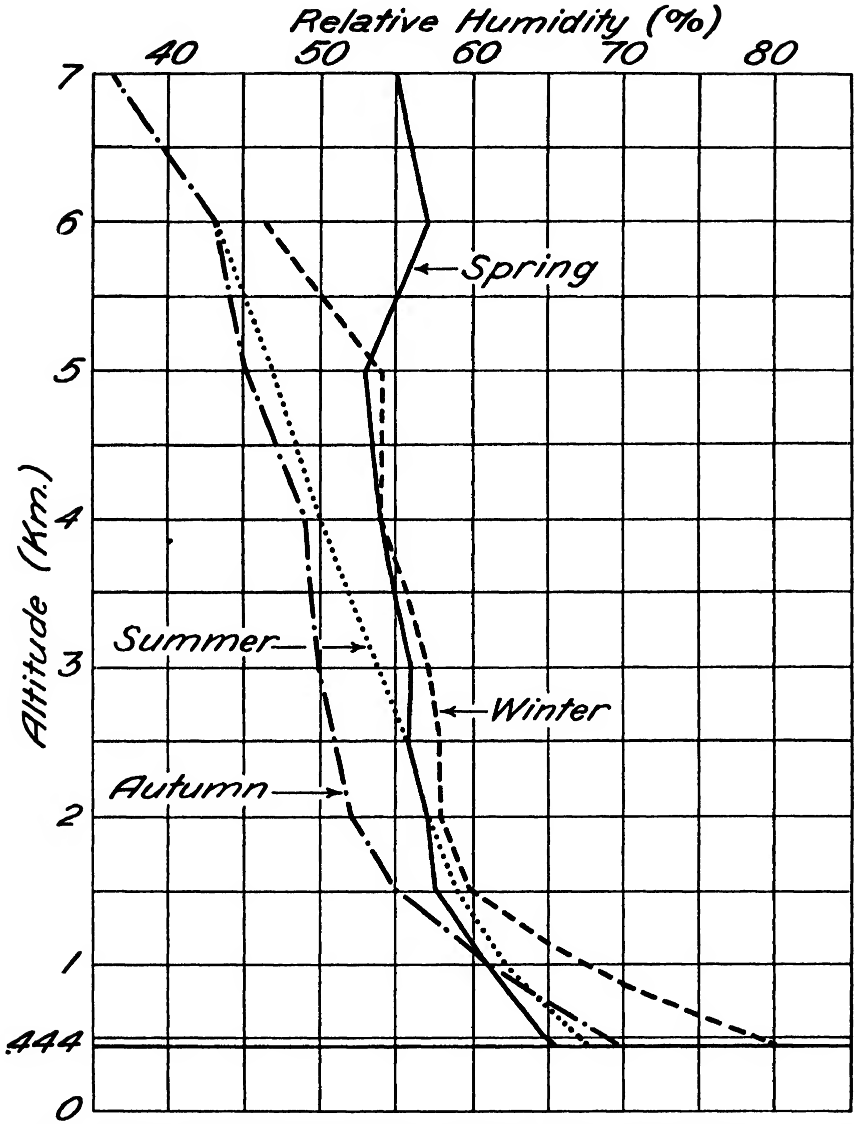

FIG. 43.—Mean seasonal free-air relative humidities (%) for Ellendale, N. Dak.

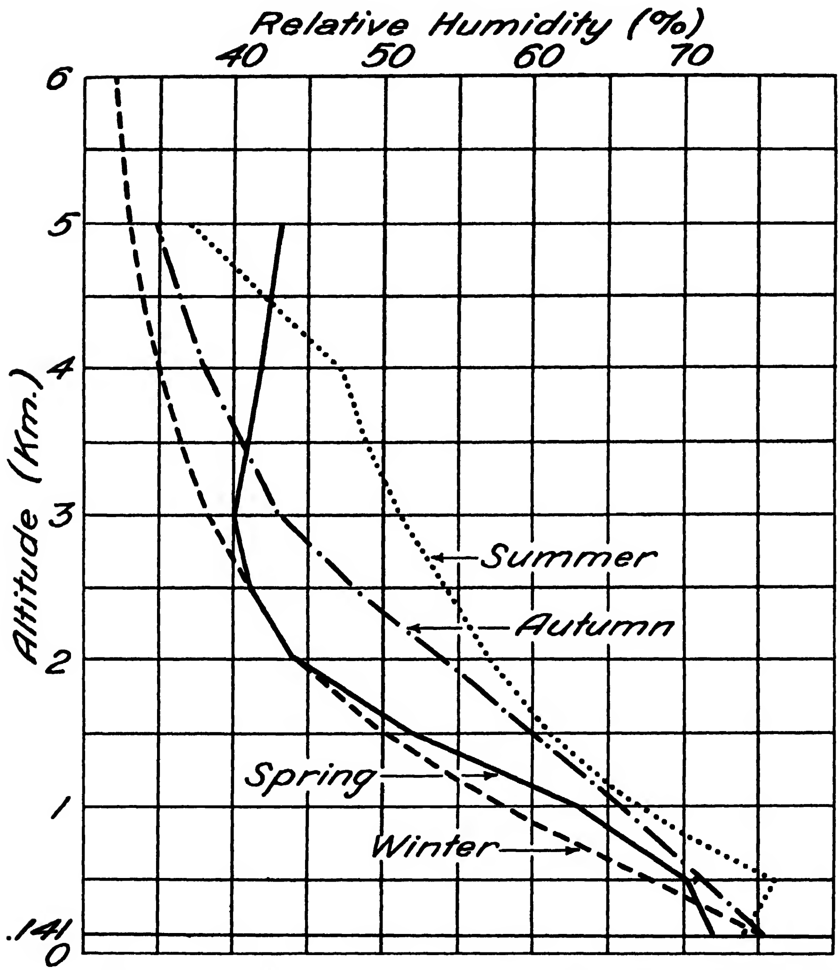

FIG. 44.—Mean seasonal free-air relative humidities (%) for Groesbeck, Texas.

4) The greatest average temperature lapse rates are found to occur at Pavia, Italy, *viz.*, 0.81° C. per 100 m. between 8 and 9 km. during summer, and 0.82° C. per 100 m. between 5 and 6 km. during winter.

Mean free-air relative humidity.—Figures 43, 44, and 45 show the mean seasonal free-air relative humidities for the same period as represented by Figures 34, 36, and 37.

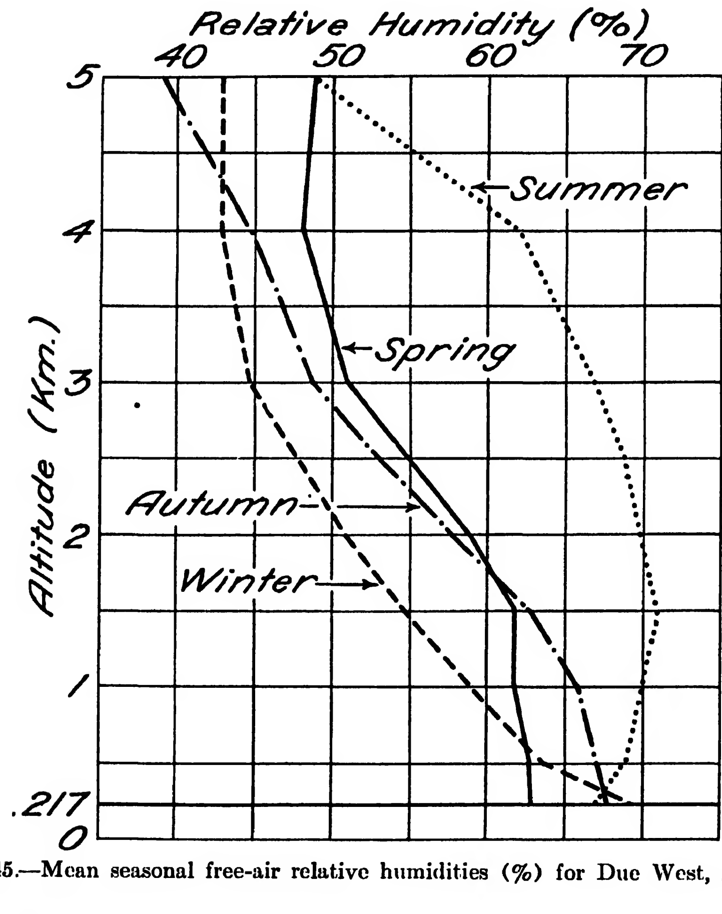

Fig. 45.—Mean seasonal free-air relative humidities (%) for Due West, S. Car.

The following features are shown:

1) The greater seasonal variation at the southern than at the northern stations, most pronounced in this respect being the high humidities of summer as compared to the other seasons at the southern stations, particularly at Due West.

2) The mean relative humidities at the higher levels during autumn, spring and winter are greater at the northern than at the southern

stations, due probably to the greater amount of storminess in the North than in the South.

Annual march of free-air vapor pressure.—Figures 46, 47, and 48

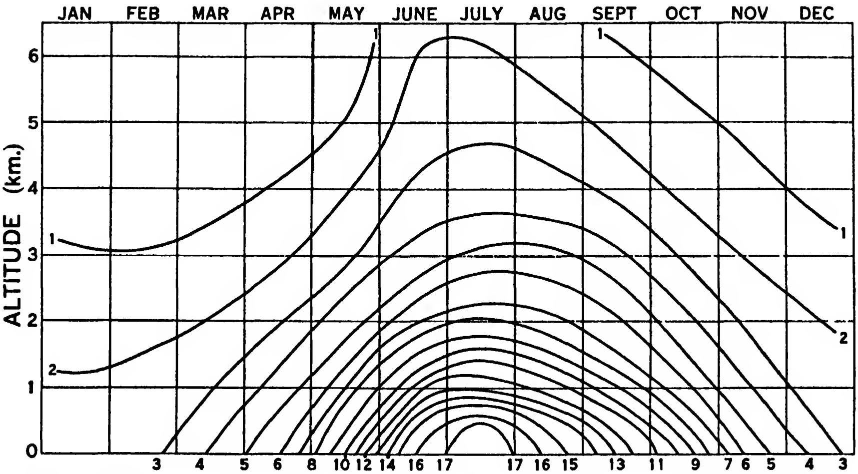

Fig. 46.—Mean annual march of free-air vapor pressures (mb.) above Ellendale, N. Dak. (Lines extended to sea-level by extrapolation.)

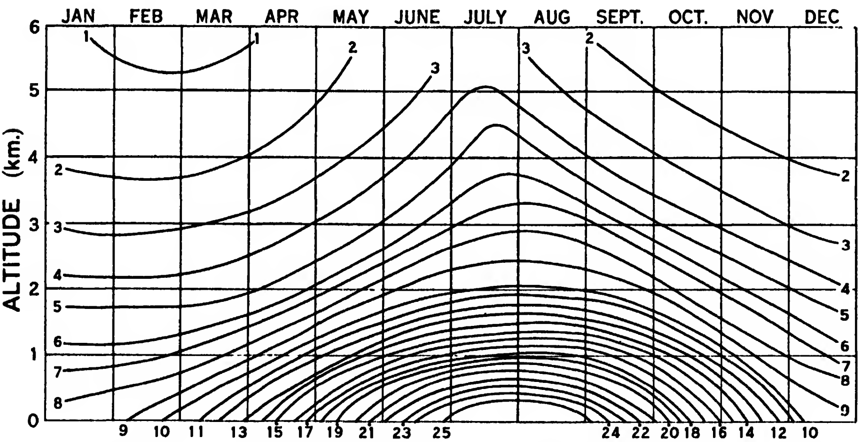

Fig. 47.—Mean annual march of free-air vapor pressures (mb.) above Groesbeck, Texas. (Lines extended to sea-level by extrapolation.)

show the features of the mean annual march of free-air vapor pressure based on kite observations for the same period as those for temperature.

The following features are noted:

1) The inverse variation of the mean vapor pressures with latitude, the differences being greater in summer than in winter.

2) The vapor pressure curves have many characteristics similar to those for temperature. A noticeable difference, however, is that the former show a smaller decrease with altitude during winter as compared with summer, especially at the northern stations. This is due to the relatively small capacity of the air for moisture when temperatures are low.

3) The most rapid rates of decrease with altitude occur at the southern stations.

4) At Due West, average vapor pressures at the upper levels during autumn are appreciably greater than at the two western stations.

Figures 49, 50 and 51 give mean seasonal free-air vapor pressures for Ellendale, Groesbeck and Due West.

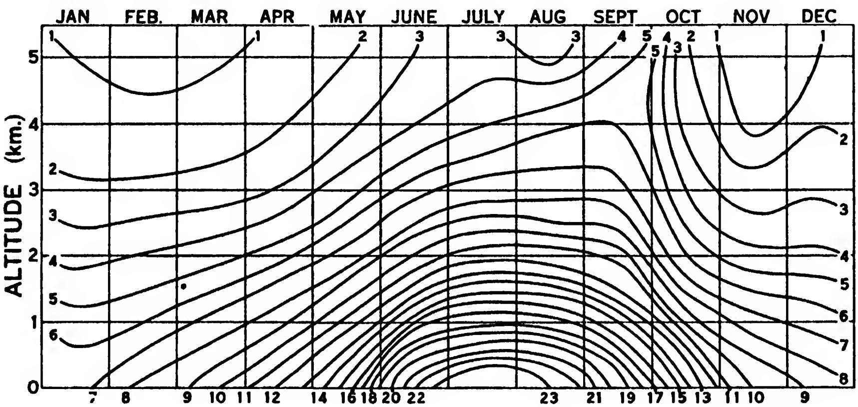

FIG. 48.—Mean annual march of free-air vapor pressures (mb.) above Due West, S. C. (Lines extended to sea-level by extrapolation).

It will be noted. 1) that the mean vapor pressures below the 3-km. level increase toward the lower latitudes in all seasons, and 2) that the seasonal relationship between the mean vapor pressures is similar to that for temperature. (See Figures 39, 40 and 41.) The curves for vapor pressure, however, converge more at high levels than do those for temperature.

Annual march of free-air resultant winds.—Figures 52-55, inclusive, show the annual march of free-air resultant wind directions and velocities for the same period as those for temperature.

The following are some of the conspicuous features shown:

1) There is a general increase in the resultant velocity with altitude, the velocities being greater at the northern than at the southern stations and greater in winter than in summer.

2) In general, the west component becomes more pronounced with increasing altitude, especially in the winter season.

3) The south component extends highest in the summer. The progressive increase in height to which this extends as the summer season advances is strikingly shown for Drexel and Groesbeck.

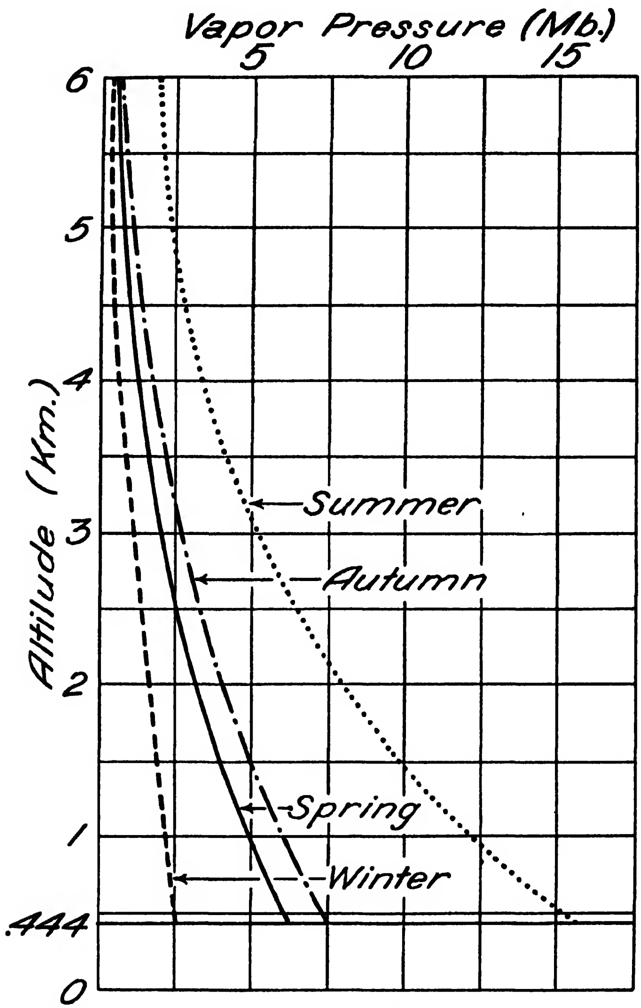

Fig. 49.—Mean seasonal free-air vapor pressures (mb.) for Ellendale, N. Dak

4) The extremely pronounced south component during the summer at Groesbeck is in marked contrast with the other stations. Conversely, the absence of a south component at Due West during the summer season is conspicuous. It will also be seen that an east component prevails at the lower levels during the autumn season at this station.

Mean free-air wind velocity.—Figure 56 shows the mean free-air wind velocities for summer and winter at Ellendale, North Dakota (Lat. 46°), the northernmost station, and Groesbeck, Texas (Lat. 31° 30′), the southernmost station, based on kite records obtained from 1918 to 1923, inclusive.

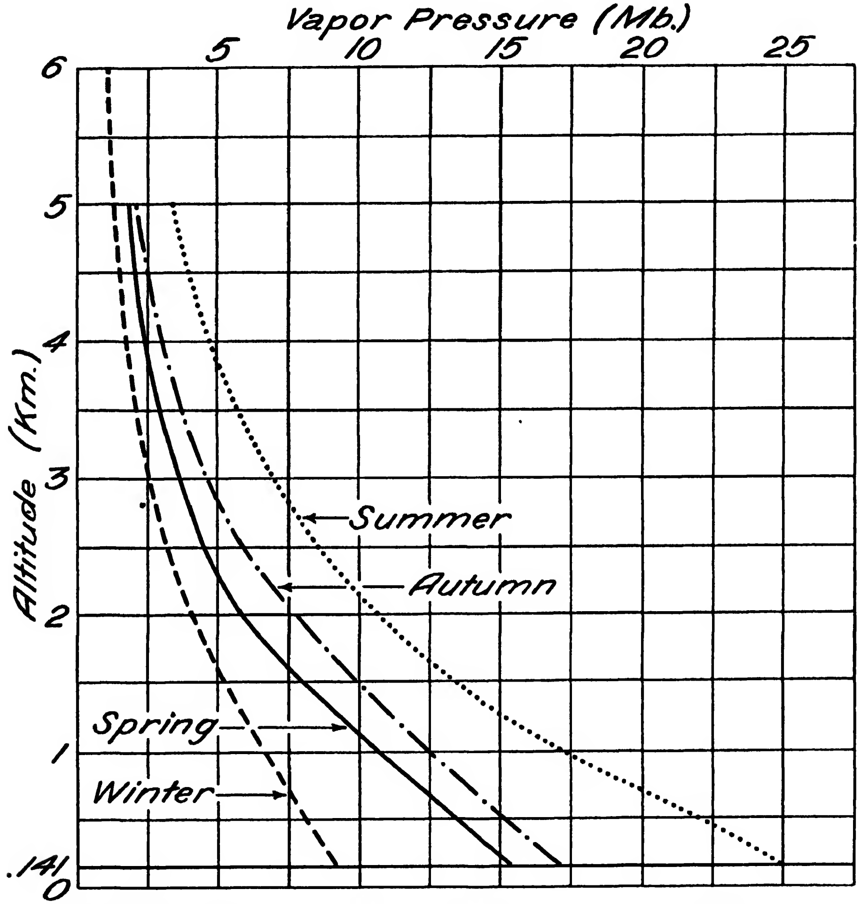

Fig. 50.—Mean seasonal free-air vapor pressures (mb.) for Groesbeck, Texas.

The mean velocities were obtained independently of wind directions. The following features are worthy of note:

1) The velocities are higher in winter than in summer at both stations.

2) The difference between the mean velocities at these stations is proportionally larger during summer than during winter. This relationship is consistent with the results found for the intermediate stations, Drexel, Nebraska (Lat. 41°), and Broken Arrow, Oklahoma (Lat. 36°), which

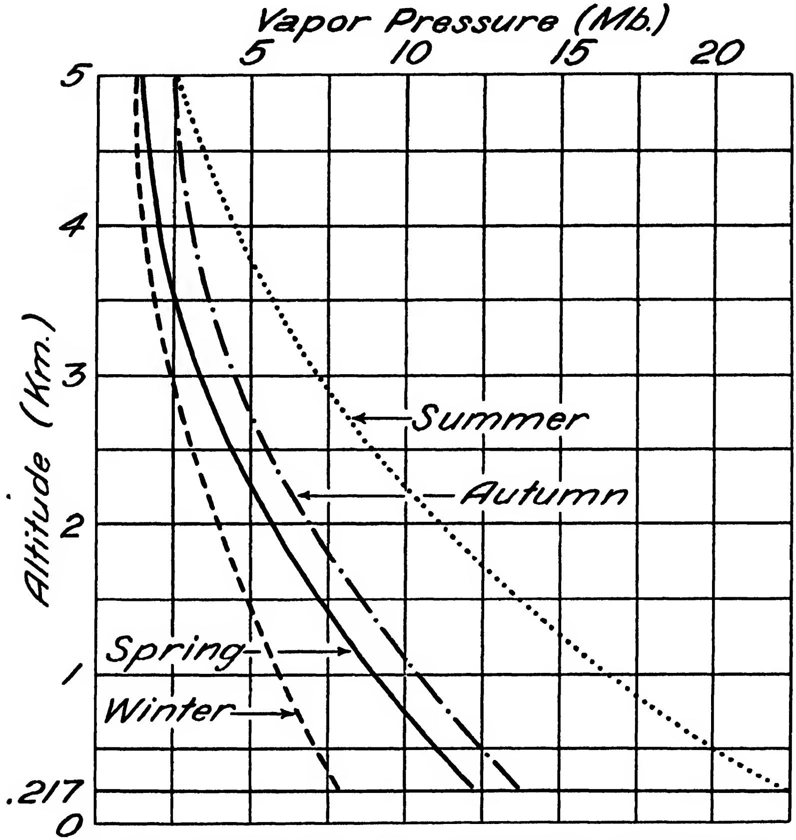

FIG. 51.—Mean seasonal free-air vapor pressures (mb.) Due West, S. C.

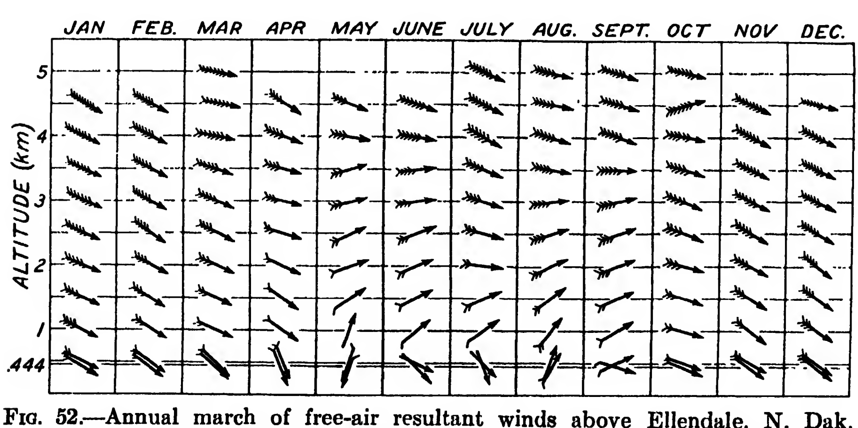

FIG. 52.—Annual march of free-air resultant winds above Ellendale, N. Dak. (Arrows fly with the wind; number of barbs represent velocity, m. p. s.)

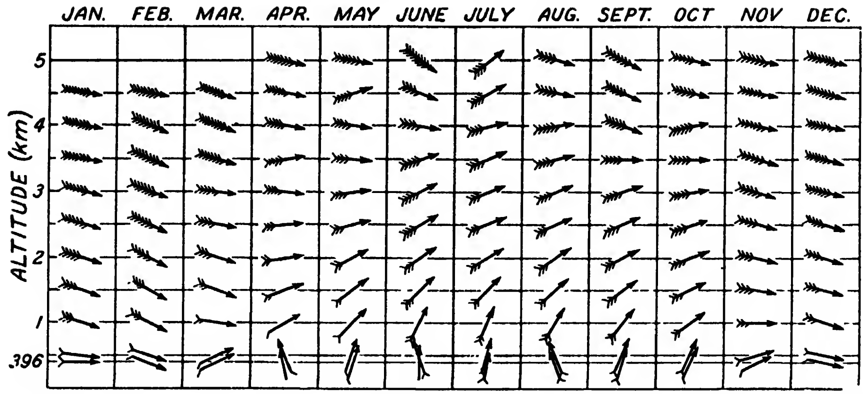

FIG. 53.—Annual march of free-air resultant winds above Drexel, Nebr. (Arrows fly with the wind; number of barbs represent velocity, m. p. s.)

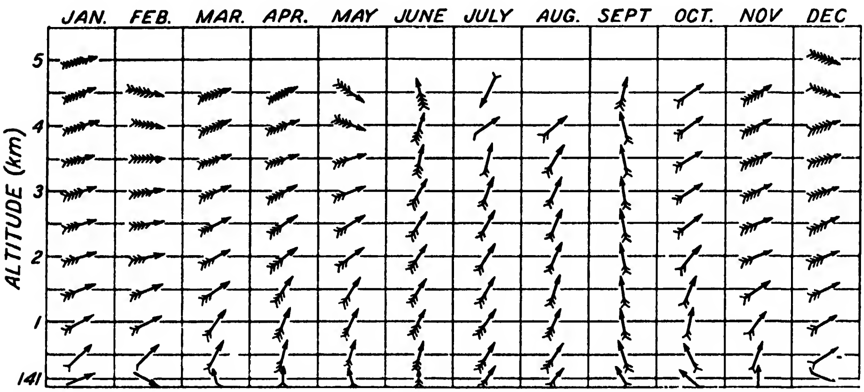

FIG. 54.—Annual march of free-air resultant winds above Groesbeck, Texas. (Arrows fly with the wind; number of barbs represent velocity, m. p. s.)

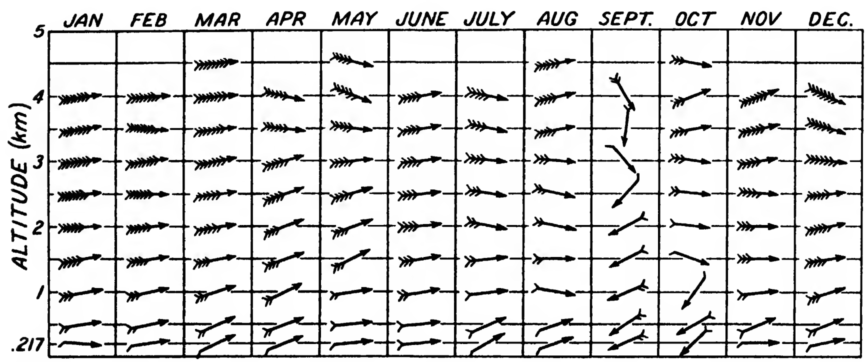

FIG. 55.—Annual march of free-air resultant winds above Due West, S. C. (Arrows fly with the wind; number of barbs represent velocity, m. p. s.)

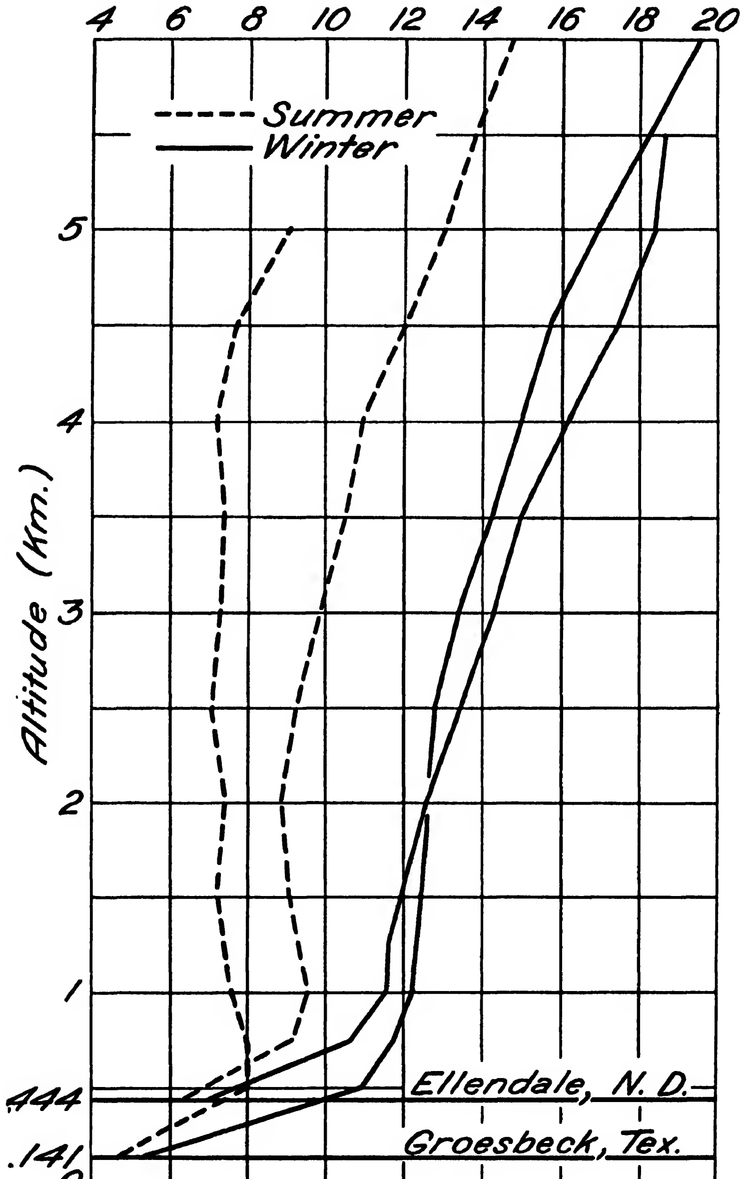

FIG. 56.—Mean free-air wind velocities (m. p. s.) at Ellendale, N. Dak. and Groesbeck, Texas.

show a much smaller latitudinal variation in the mean free-air velocities during winter than during summer.[29] This is due to the fact that the permanent belt of high pressure and light winds extends north during summer to the latitude of Groesbeck, whereas in winter the stronger latitudinal temperature gradients cause relatively strong winds over this region.

3) The mean velocities in winter are actually greater at Groesbeck than at Ellendale at corresponding altitudes above sea-level up to 2,000 meters.

4) A comparatively large increase in velocity with altitude occurs in the first few hundred meters above the surface. At Groesbeck this condition extends to a greater height above the surface than at Ellendale.

5) Superimposed upon the ground stratum wherein the velocity increases rapidly with height there is a layer characterized by a very small increase during the winter and a slight decrease during the summer. Above the 4-km. level the average velocity again increases with elevation. A decrease in velocity with altitude again occurs above the 12-km. level. This is strikingly brought out in Figure 57 which shows the daily wind velocity curves (m. p. s.), as obtained from daily double-theodolite observations at Groesbeck, Texas, during October, 1927.[30] The consistent decrease in wind velocity in the stratosphere of these individual observations is very striking.

Mean free-air wind direction.—Figures 58 to 61, inclusive, show the mean free-air wind directions as related to surface directions for summer and winter at Ellendale, North Dakota, and Groesbeck, Texas, based on kite records obtained from 1918 to 1923, inclusive.

The following are some of the chief features shown:

1) The turning of winds with altitude is greater in winter than in summer and the turning at Ellendale is greater than at Groesbeck in both seasons. Since the wind direction is determined by the pressure gradient, which in turn is a result of the horizontal temperature gradient, it follows that winds will become westerly at a lower altitude when the horizontal temperature gradient, *i. e.*, south to north, is comparatively steep. Horizontal temperature gradients are ordinarily steeper in winter than in summer and steeper at northern than at southern stations.

2) The average turning with altitude of surface winds between NW and NNE is to the left; of those between NE, SE, and SW, to the right, during winter at both stations, and during the summer at Ellendale. During the summer at Groesbeck the surface winds of all directions show practically no turning with altitude to the heights shown by the observations. A slight veering with altitude, however, occurs for a short distance above the surface with practically all winds, owing to the friction

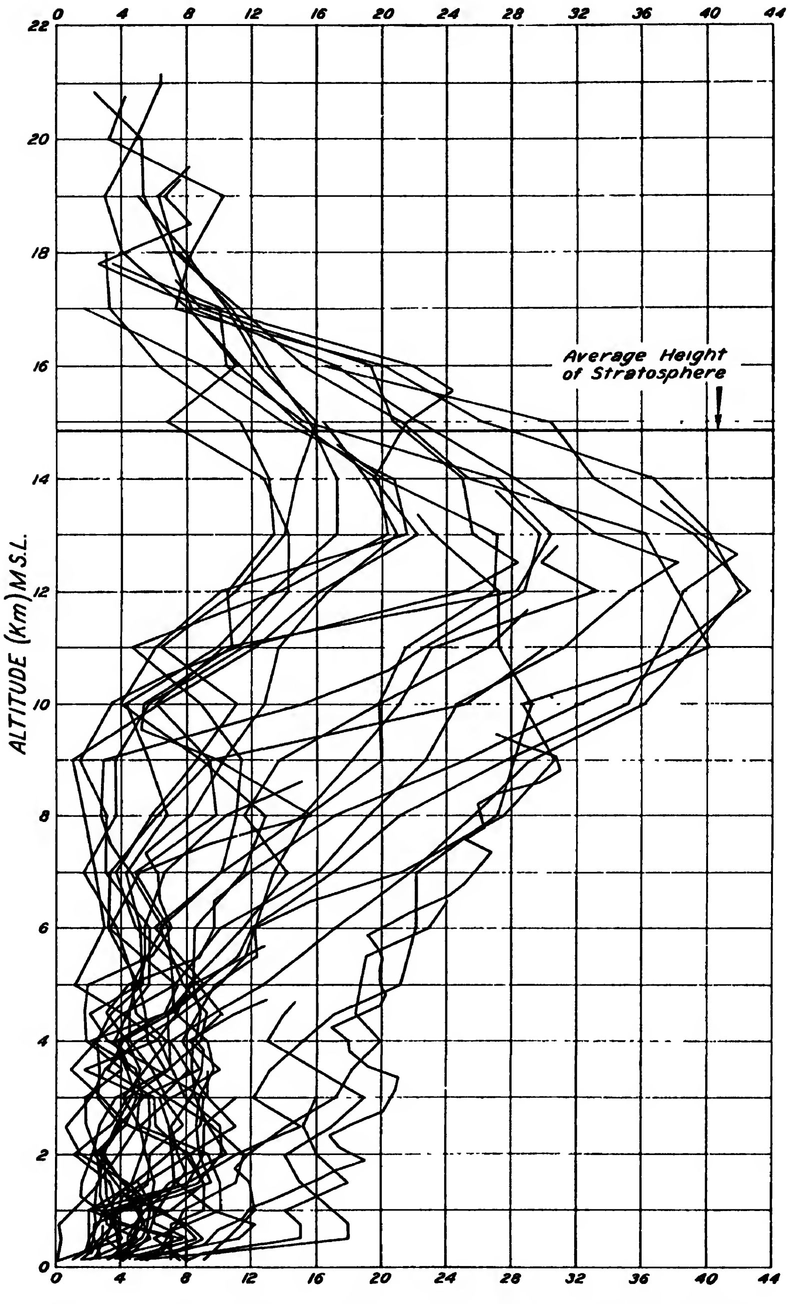

FIG. 57.—Daily wind velocity curves (m. p. s.) Groesbeck, Texas, October 1927.

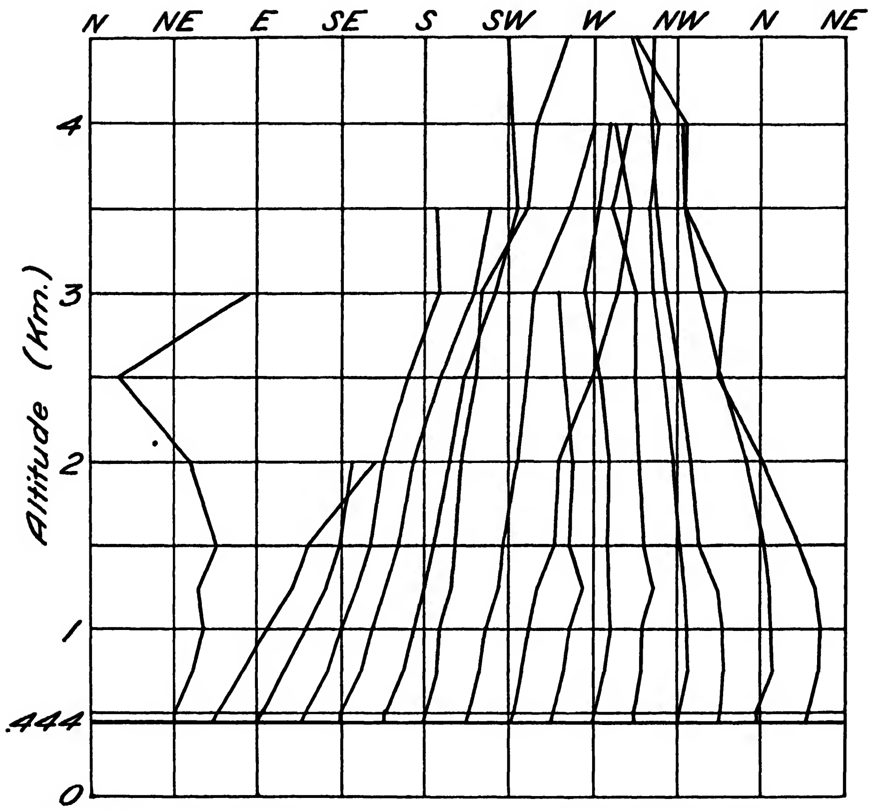

FIG. 58.—Mean free-air wind directions as related to surface directions for summer at Ellendale, N. Dak.

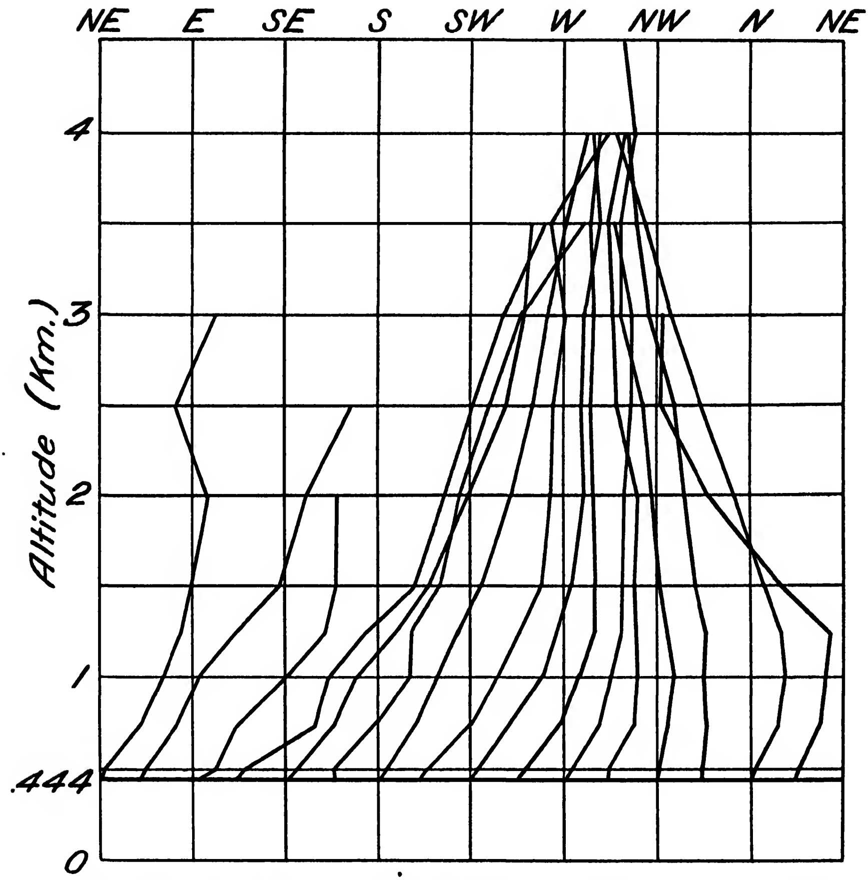

FIG. 59.—Mean free-air wind directions as related to surface directions for winter at Ellendale, N. Dak.

of the air over the ground. This friction causes the air to flow from higher to lower pressure instead of parallel to the isobars as is the case in the free air.

It has been shown that in the United States a north or south component in the winds at and near the surface persists in a majority of cases at all levels in the troposphere, and presumably well up into the stratosphere,

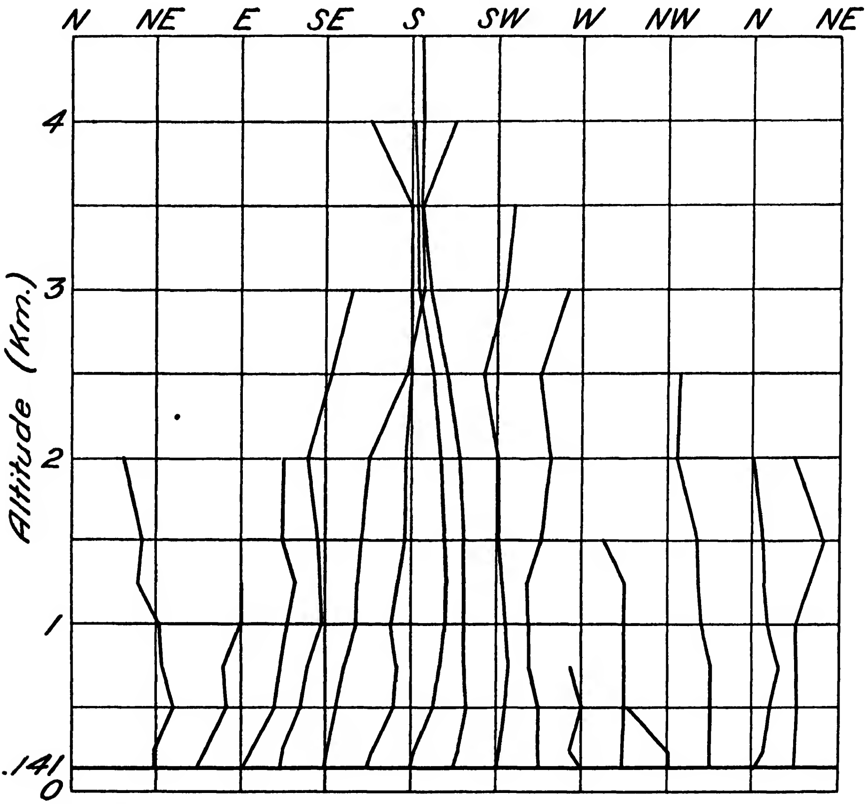

FIG. 60.—Mean free-air wind directions as related to surface directions for summer at Groesbeck, Texas.

although the wind direction itself usually changes, *e. g.*, NNW backing to NW or WNW; SE veering to SW or WSW.[31]

The changes in wind direction with elevation to altitudes well within the stratosphere are shown in Figure 62 and are based on a series of sounding balloon observations (double-theodolite) made at Groesbeck, Texas, in October 1927.[30] These results are representative of this latitude (31° 30′ N.) and season.

The more conspicuous features are:

1) A small change in direction between the surface and 10 to 12 km.
2) A veering or backing to westerly occurs at about 12 km.
3) A consistent westerly direction between 12 and 17 km.
4) A veering to easterly above 18 km.

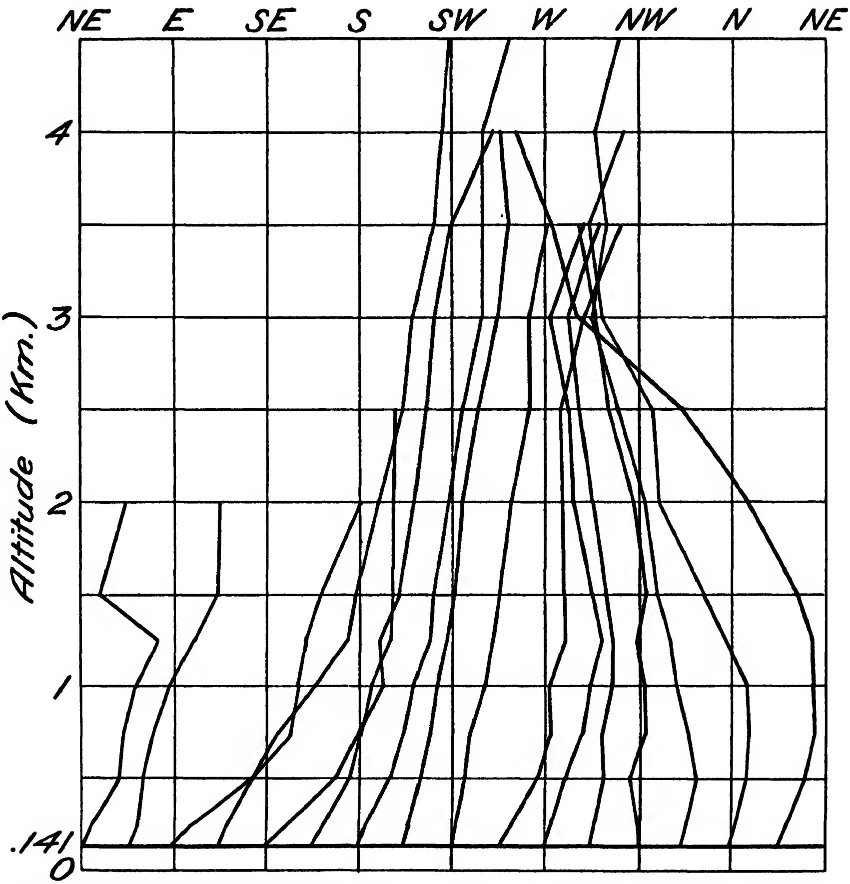

FIG. 61.—Mean free-air wind directions as related to surface directions for winter at Groesbeck, Texas.

Diurnal march of free-air winds.—Figures 63, 64 and 65 show the average a. m. and p. m. wind velocities at Ellendale, North Dakota, Groesbeck, Texas, and Key West, Florida, respectively.[32] It will be seen that the diurnal variation above the 100-meter level is of opposite phase to that at the surface. In other words, instead of the maximum average velocity occurring in the afternoon as at the surface, it occurs in the early morning in the free air. This results in a much more rapid increase in

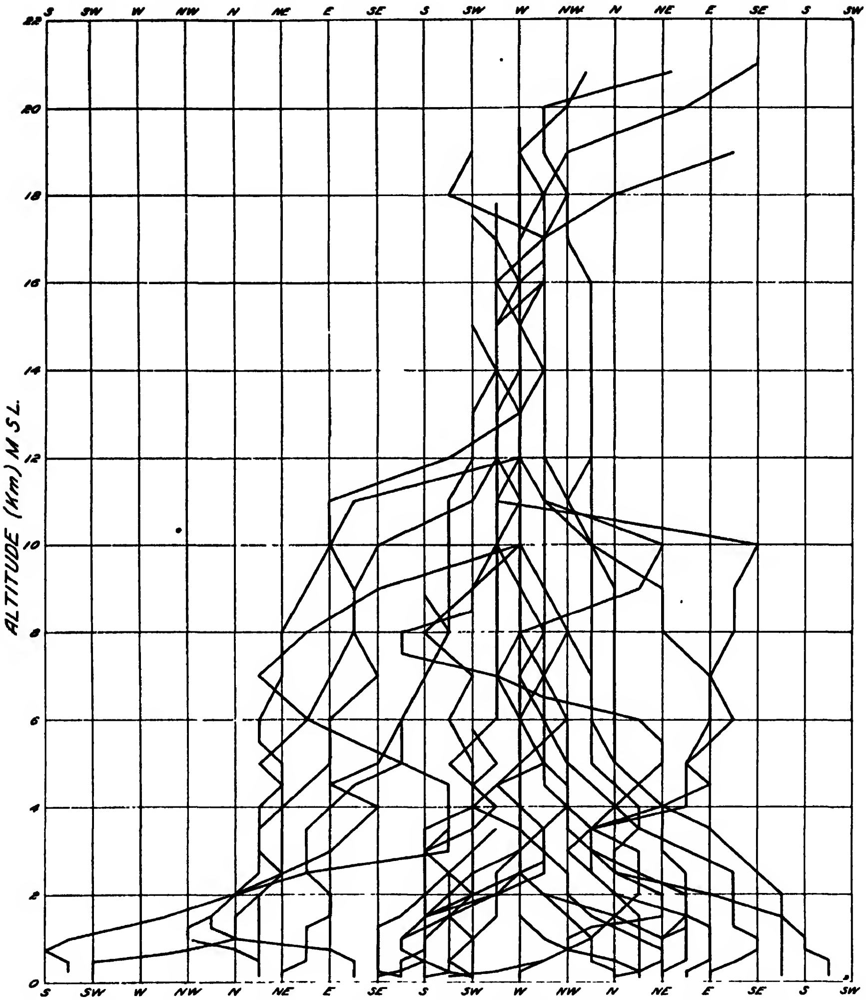

FIG. 62.—Daily wind direction curves, Groesbeck, Texas, October 1927.

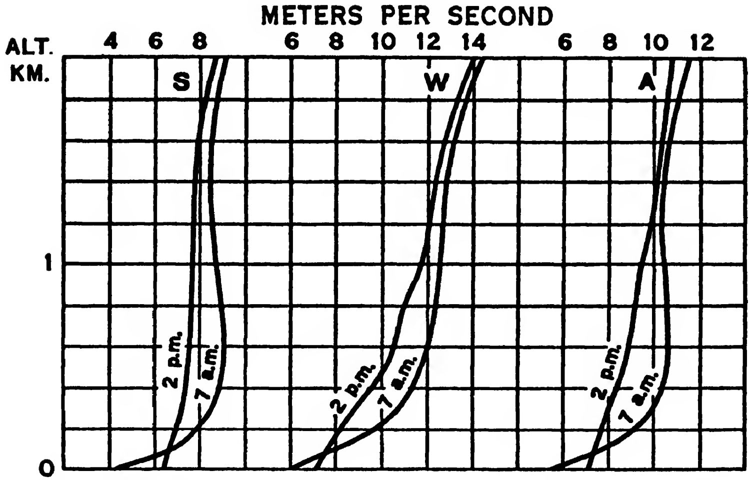

FIG. 63.—Average a. m. and p. m. wind velocities at Ellendale, N. Dak. S = summer, W = winter, A = annual.

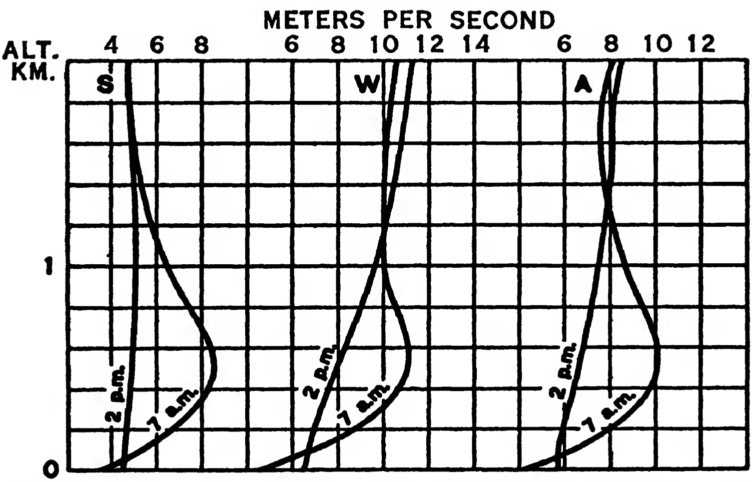

FIG. 64.—Average a. m. and p. m. wind velocities at Groesbeck, Texas. S = summer, W = winter, A = annual.

velocity throughout the first few hundred meters in the morning than during the afternoon, as is strikingly shown in the figures. In this connection the following statement is quoted from the Monthly Weather Review Supplement No. 26:

The diurnal range at the surface is approximately 1 to 2 m.p.s. The change of phase occurs between 50 and 100 meters. Above this level there is a rapid increase in the range to about 500 meters, where it amounts to 2 to 4 m.p.s., except that at Key West it is only about 1 m.p.s. The diurnal variation practically ceases at 1,500 to 2,000 meters, the height of no change being apparently somewhat higher in summer than in winter. There does not seem to be any marked seasonal variation in the range at about 500 meters, nor is there any variation that can be attributed to latitude alone. There is, however, a some-

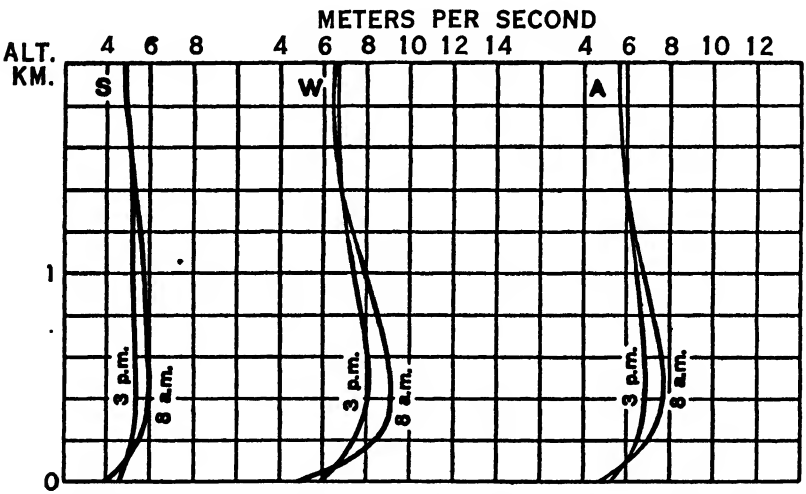

FIG. 65.—Average a. m. and p. m. wind velocities at Key West, Fla. S = summer, W = winter, A = annual.

what greater range at those stations where the diurnal temperature range is large and in consequence convection is active, *e. g.*, Drexel, Broken Arrow and Groesbeck. On the other hand, at Key West, far removed from any extended land areas and relatively free from appreciable convectional activity, the range both at the surface and above is very small, and ceases altogether at about 1,400 meters. Moreover, near the surface the change with altitude in the afternoon is similar to that in the morning, though less pronounced, whereas at the other stations the character of the a.m. and p.m. curves is distinctly different.

Mean free-air temperatures for different surface wind directions.—In Figures 66 and 67 are shown the mean free-air temperatures as related to surface wind directions during summer and winter at Ellendale, North Dakota, and Groesbeck, Texas, based on kite records obtained from 1918 to 1921, inclusive.

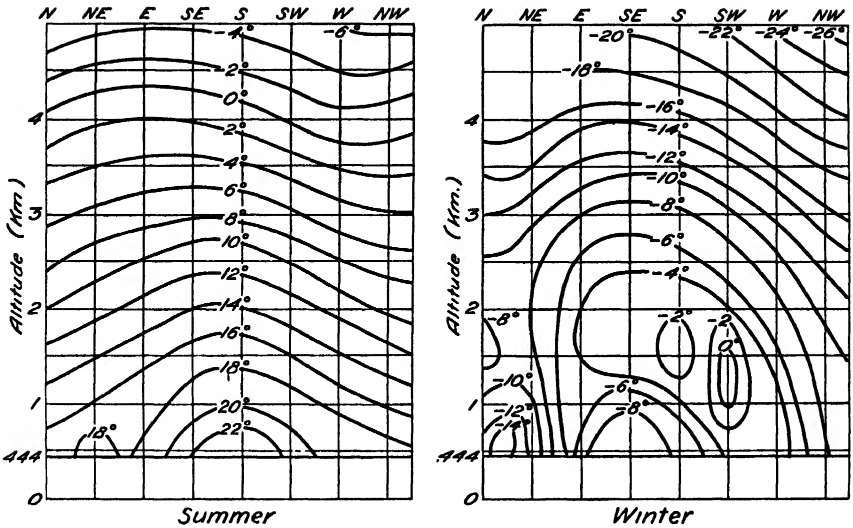

Fig. 66.—Mean free-air temperatures (°C.) as related to surface wind directions, Ellendale, N. Dak.

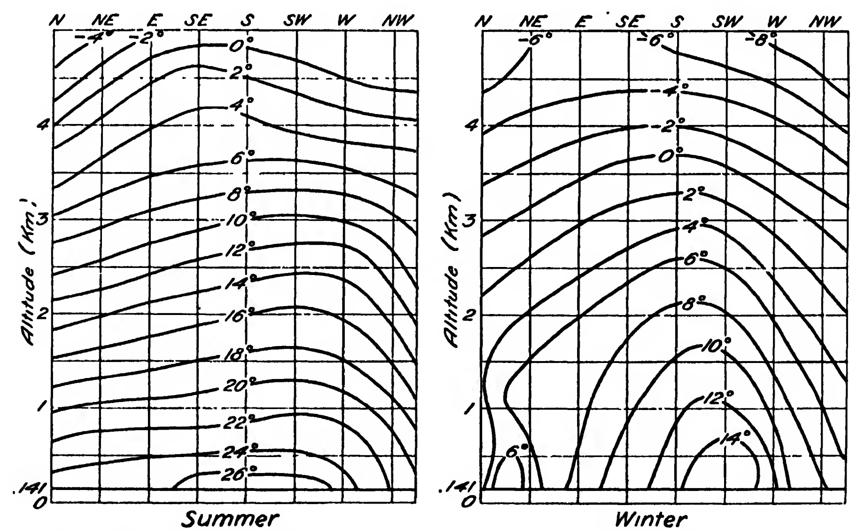

Fig. 67.—Mean free-air temperatures (°C.) as related to surface wind directions, Groesbeck, Texas.

The following are the chief features shown:

1) There is greater persistence of higher temperatures with southerly than with northerly surface winds, this relationship being more pronounced in winter than in summer.

2) The highest average temperatures at the surface and at lower levels occur with southwesterly surface winds, but with increasing altitude they occur with a backing of the surface wind, so that at 4 and 5 km. the highest temperatures are found with easterly surface winds. This is due to the fact that the south component usually increases with altitude with easterly surface winds.

In a study of the relations between free-air temperatures and wind directions [31] the following have been found:

1) Temperatures accompanying southerly winds are, on the average, considerably higher than those accompanying northerly winds at all levels in the troposphere.

2) The difference is more pronounced at 1 and 2 km. than at greater heights or at the surface; at 3 and 4 km. it is essentially the same as at the surface; above 4 km. it gradually diminishes, becoming zero at the upper limit of the troposphere. In the stratosphere the reverse relation is found, viz., lower temperatures with south than with north component winds.

3) The relations given in 1) and 2) are more pronounced in winter than in summer and at northern than at southern stations, *i. e.*, when and where the latitudinal temperature gradient is strongest. Exceptions to these relations are due either to a temporary reversal in the normal latitudinal distribution of temperature (occasional summer condition) or to the importation of large masses of cold or warm air in rapidly moving HIGHS or LOWS or to the fact that in some instances the wind direction does not represent the original source of the air, the latter having followed a curved path round a nearly stationary HIGH or LOW.

4) Changes from north to south component winds are in nearly all cases accompanied by rising temperatures and vice versa.

5) Owing to effects of temperature on air density, the free-air position of a LOW is usually to the northwest of its sea-level position, and that of a HIGH to the southwest. Winds therefore are southwesterly above the sea-level positions of LOWS and northwesterly above the sea-level positions of HIGHS. Under these conditions the air above LOWS is on the average warmer than that above HIGHS, the effects of importation being much greater than those of vertical movement.

6) When easterly winds prevail from the surface up to 3 or 4 kilometers, HIGHS and LOWS are either stationary or their movements

are slow and erratic. Such HIGHS and LOWS are nearly circular and probably symmetrical to great heights; air circulation in them is fairly definite and steady, and the effects of vertical movement of the air are greater than those of importation, the centers of HIGHS being warmer than the centers of LOWS.

7) Since in this country symmetrical HIGHS and LOWS, referred to in 6), are less frequent than those with a westward shift of the centers, referred to in 5), it follows that the air above the sea-level positions of HIGHS is on the average colder than that above the sea-level positions of LOWS. If, however, we take the lowest and highest pressures at different heights as the basis of comparison, we find that the lowest pressures are accompanied by the lowest temperatures, the pressure itself at any level being largely a function of the mean temperature of the air column beneath.

Mean free-air barometric and vapor pressures, temperatures and densities.—These data are given in Table 8 and graphically in Figures 68, 69, and 70. They have been adopted by the National Advisory Committee for Aeronautics as the "standard atmosphere,"[28] and are representative of average conditions at latitude 40° N. in the United States. Therefore, with increasing distance from latitude 40°, the variation in the mean values also increases. The variation in temperature is greatest in winter, when it amounts to about 1° C. per degree of latitude at the surface, diminishing slightly at higher levels.

The standard atmosphere.—With the advance of aeronautics and the science of artillery, engineers and other specialists in these fields have come to require a specific knowledge of the varying states of the atmosphere from the ground up to very great heights. This has led to the introduction of a conventional term commonly known as the "standard atmosphere," which pretends to specify the normal or average condition. As is well known, the "standard atmosphere" is never found; that is to say, at no time or place do "standard" or average conditions of all of the meteorological elements at all altitudes simultaneously occur. Nevertheless it is proper, and in certain fields (especially those of aviation and ordnance) it is necessary, to adopt so-called "standard" values, and it is desirable to have these represent as closely as possible true mean values in order that corrections due to departures from these means may be comparatively small in most cases. Hence, the adoption of an "isothermal atmosphere," proposed by some investigators, although a desirable simplification in some respects, is inadvisable because of the large corrections that would have to be applied at practically all altitudes. Although a knowledge of temperature may not be vital in aerodynamic

TABLE 8

MEAN FREE-AIR BAROMETRIC AND VAPOR PRESSURES, TEMPERATURES AND DENSITIES AT ABOUT LATITUDE 40° IN THE UNITED STATES

Altitude, mean sea level	Pressure		Temperature		Vapor pressure		Density: Per cent standard	Density: Kilograms per cubic meter
				Summer				
m.	*mb.*	*mm.*	*° C.*	*° A.*	*mb.*	*mm.*		
0	1,014.0	760.5	25.0	298.0	22.0	16.5	90.9	1.175
500	957.5	718.0	22.0	295.0	17.5	13.0	86.8	1.123
1,000	904.0	678.0	19.0	292.0	14.0	10.5	82.9	1.072
1,500	852.5	639.5	16.0	289.0	11.0	8.5	79.1	1.023
2,000	803.5	602.5	13.0	286.0	8.5	6.5	75.4	0.975
2,500	757.0	568.0	10.0	283.0	6.5	5.0	71.8	.929
3,000	713.0	535.0	7.0	280.0	5.0	4.0	68.4	.885
4,000	630.5	473.0	0.5	273.5	3.5	2.5	62.0	.801
5,000	556.0	417.0	− 5.5	267.5	2.0	1.5	55.9	.723
6,000	488.5	366.5	− 12.5	260.5	1.0	1.0	50.5	.653
7,000	428.0	321.0	− 19.5	253.5	0.5	0.5	45.4	.587
8,000	373.5	280.0	− 26.0	247.0	.....	.....	40.7	.527
9,000	324.5	243.5	− 32.5	240.5	.....	.....	36.4	.470
10,000	281.0	211.0	− 39.0	234.0	.....	.....	32.4	.418
11,000	242.5	182.0	− 45.5	227.5	.....	.....	28.7	.371
12,000	208.5	156.5	− 52.0	221.0	.....	.....	25.4	.329
13,000	178.5	134.0	− 55.0	218.0	.....	.....	22.1	.285
14,000	152.5	114.5	− 55.0	218.0	.....	.....	18.8	.244
15,000	130.5	98.0	− 55.0	218.0	.....	.....	16.1	.209
16,000	111.5	83.5	− 55.0	218.0	.....	.....	13.8	.178
17,000	95.5	71.5	− 55.0	218.0	.....	.....	11.8	.153
18,000	81.5	61.0	− 55.0	218.0	.....	.....	10.1	.130
19,000	69.5	52.0	− 55.0	218.0	.....	.....	8.6	.111
20,000	59.5	44.5	− 55.0	218.0	.....	.....	7.4	.095
				Winter				
0	1,020.0	765.0	− 2.0	271.0	4.5	3.5	101.2	1.309
500	957.5	718.0	− 3.0	270.0	3.5	2.5	95.4	1.234
1,000	899.0	674.5	− 3.0	270.0	3.0	2.5	89.6	1.159
1,500	844.0	633.0	− 4.0	269.0	2.5	2.0	84.4	1.092
2,000	792.0	594.0	− 5.0	268.0	2.0	1.5	79.6	1.029
2,500	743.0	557.5	− 7.0	266.0	2.0	1.5	75.2	0.972
3,000	697.0	523.0	− 9.0	264.0	1.5	1.0	71.0	.918
4,000	611.5	458.5	− 14.5	258.5	1.0	1.0	63.7	.823
5,000	535.0	401.5	− 20.5	252.5	0.5	0.5	57.1	.738
6,000	466.5	350.0	− 27.5	245.5	.....	.....	51.2	.662
7,000	405.5	304.0	− 34.5	238.5	.....	.....	45.8	.592
8,000	350.5	263.0	− 41.0	232.0	.....	.....	40.7	.526
9,000	302.0	226.5	− 46.5	226.5	.....	.....	35.9	.464
10,000	259.5	194.5	− 50.0	223.0	.....	.....	31.4	.405

TABLE 8—*Continued*

Altitude, mean sea level	Pressure		Temperature		Vapor pressure		Density: Per cent standard	Density: Kilograms per cubic meter
Winter—Continued								
11,000	222.5	167.0	−52.5	220.5	.....	.....	27.2	.352
12,000	190.5	143.0	−54.0	219.0	.....	.....	23.4	.303
13,000	163.0	122.5	−55.0	218.0	.....	.....	20.1	.260
14,000	139.5	104.5	−55.0	218.0	.....	.....	17.2	.223
15,000	119.0	89.5	−55.0	218.0	.....	.....	14.7	.190
16,000	102.0	76.5	−55.0	218.0	.....	.....	12.6	.163
17,000	87.0	65.5	−55.0	218.0	.....	.....	10.8	.139
18,000	74.5	56.0	−55.0	218.0	.....	.....	9.2	.119
19,000	64.0	48.0	−55.0	218.0	.....	.....	7.9	.102
20,000	54.5	41.0	−55.0	218.0	.....	.....	6.7	.087
Annual *								
m.	*mb.*	*mm.*	° *C.*	° *A.*	*mb.*	*mm.*		
0	1,017.0	763.0	11.5	284.5	11.5	8.5	95.9	1.240
500	957.5	718.0	9.5	282.5	9.5	7.0	91.0	1.176
1,000	901.5	676.0	8.0	281.0	7.5	5.5	86.2	1.114
1,500	848.5	636.5	6.0	279.0	6.0	4.5	81.7	1.056
2,000	798.0	598.5	4.0	277.0	5.0	4.0	77.4	1.001
2,500	750.5	563.0	1.5	274.5	4.0	3.0	73.5	0.951
3,000	705.0	529.0	− 1.0	272.0	3.0	2.5	69.7	.902
4,000	621.0	466.0	− 7.0	266.0	2.0	1.5	62.8	.813
5,000	546.0	409.5	−13.0	260.0	1.0	1.0	56.5	.731
6,000	478.0	358.5	−20.0	253.0	0.5	0.5	50.8	.657
7,000	417.0	313.0	−27.0	240.0	.....	.....	45.7	.591
8,000	362.0	271.5	−33.5	239.5	.....	.....	40.7	.527
9,000	313.5	235.0	−39.5	233.5	.....	.....	36.2	.468
10,000	270.5	203.0	−44.5	228.5	.....	.....	31.9	.412
11,000	232.5	174.5	−49.0	224.0	.....	.....	28.0	.362
12,000	199.5	149.5	−53.0	220.0	.....	.....	24.4	.316
13,000	171.0	128.5	−55.0	218.0	.....	.....	21.1	.273
14,000	146.0	109.5	−55.0	218.0	.....	.....	18.0	.233
15,000	125.0	94.0	−55.0	218.0	.....	.....	15.4	.200
16,000	107.0	80.5	−55.0	218.0	.....	.....	13.2	.171
17,000	91.5	68.5	−55.0	218.0	.....	.....	11.3	.146
18,000	78.0	58.5	−55.0	218.0	.....	.....	9.6	.125
19,000	67.0	50.5	−55.0	218.0	.....	.....	8.3	.107
20,000	57.0	43.0	−55.0	218.0	.....	.....	7.0	.091

* The annual means also represent quite closely the average spring and autumn conditions.

tests, it certainly is important when the thermodynamic or power production phase is considered. Moreover, in the design, construction and use of altimeters a knowledge of the altitude-pressure relation is essential, and this relation varies decidedly with temperature. What is needed, then, in defining the "standard atmosphere," is a series of values of

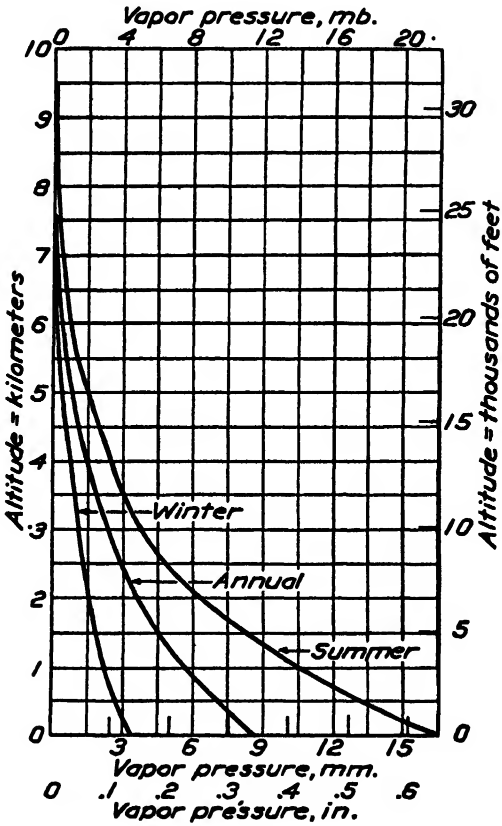

FIG. 68.—Mean free-air vapor pressures at about latitude 40° N. in the United States.

pressure, temperature, and density, at different altitudes, which represent as closely as possible actual average conditions. If tables or curves were prepared for different places and seasons, the corrections for variations from standard or average values would in each case be comparatively small and easily applied. Such a procedure would, however, complicate the matter, since it would necessitate the use of a large number of tables and would make impossible the comparison of tests at different places.

It seems desirable, therefore, to select data for some place or places so located that the results shall be as nearly as possible representative of conditions in the entire region in which they will be used.

Mean free-air wind velocities in HIGHS and LOWS.—Figures 71 to 74, inclusive, show the mean free-air wind velocities during winter in the

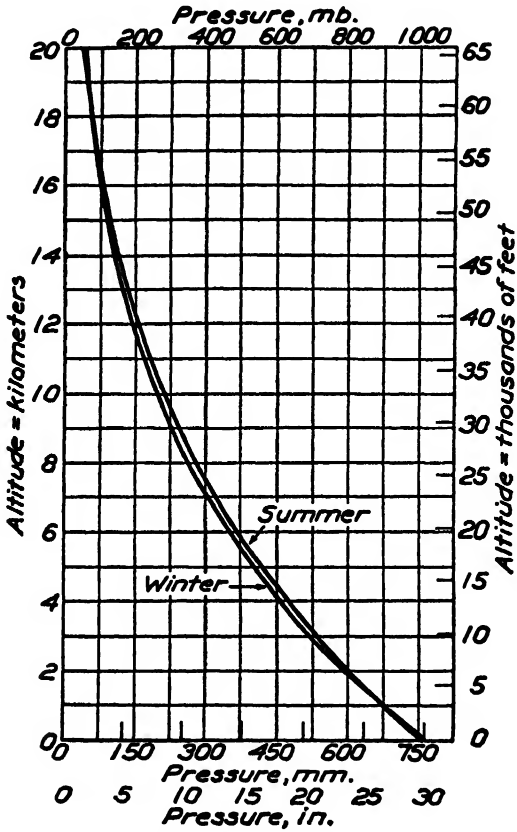

Fig. 69.—Mean free-air barometric pressures at about latitude 40° N. in the United States.

various quadrants of well pronounced anticyclones (HIGHS) and cyclones (LOWS) in the United States for the region east of the Rocky Mountains.[33] The quadrant of the pressure area and the number of observations averaged are indicated at the top of each curve. The data for the winter season were chosen because at that time cyclones and anticyclones are most highly developed and their distinguishing characteristics most pronounced.

It will be noted that 1) the strongest winds occur in the southern sector, *i. e.,* the SW and SE quadrants of LOWS, whereas in HIGHS, they are found in the eastern sector, *i. e.,* the SE and NE quadrants.

2) The lightest winds occur in the NW quadrant of LOWS and in the SW quadrant of HIGHS.

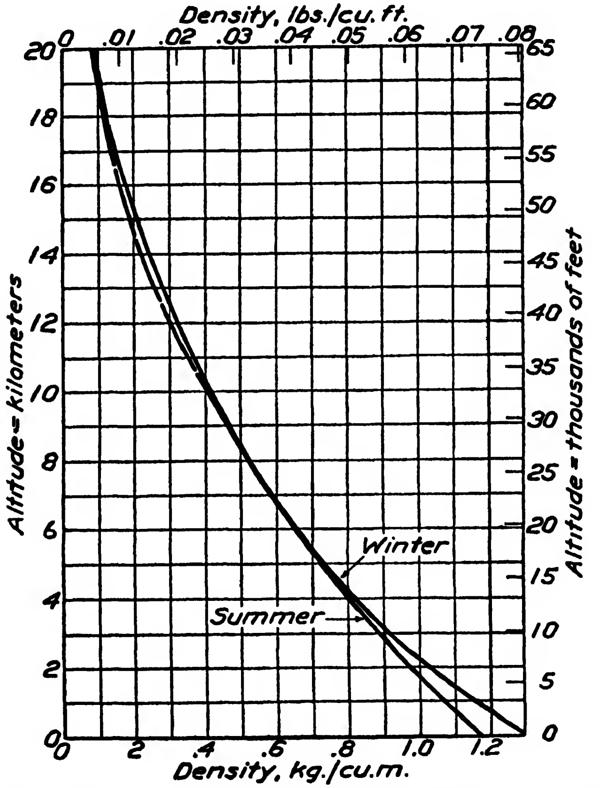

FIG. 70.—Mean free-air densities at about latitude 40° N. in the United States.

3) The winds in the lower levels, *i. e.*, the first 500 m., average appreciably greater in LOWS than in HIGHS. Above that elevation, there is a very marked decrease in the NE and NW quadrants and center of LOWS and in the SW quadrant of HIGHS.

4) The mean wind velocities are very light in the lower levels of the central region of HIGHS, being lighter than in any of the four quadrants. In the upper levels of the central region, however, they increase considerably, and conform closely to the mean values of the eastern sector, *i. e.*, the region where upper winds are strongest.

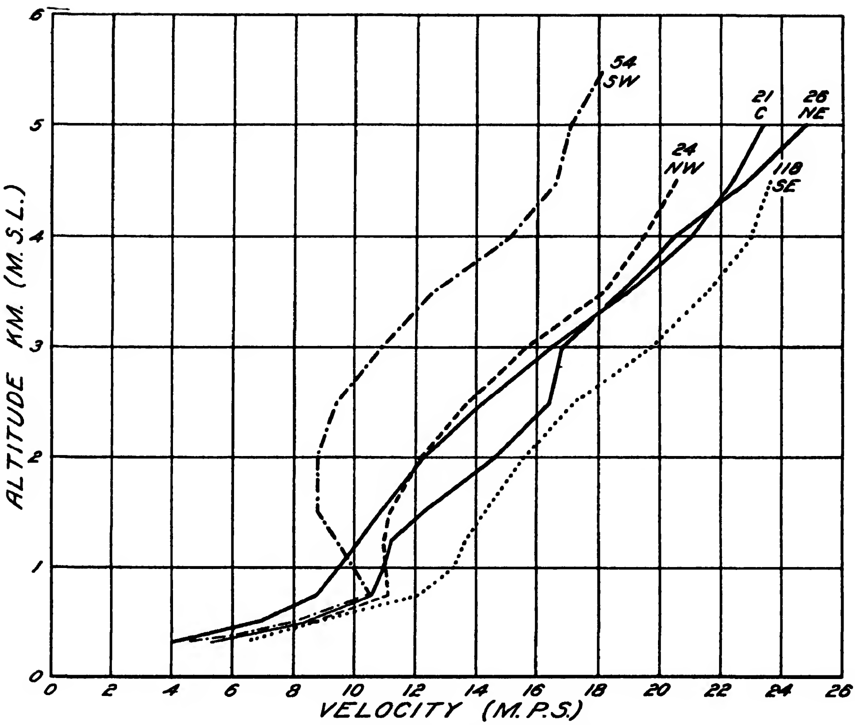

Fig. 71.—Anticyclonic wind velocity, winter, northern United States.

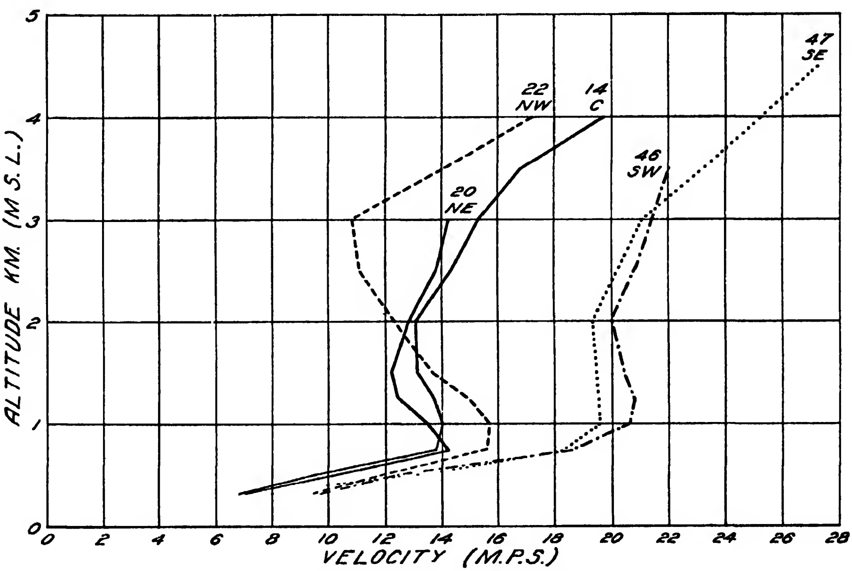

Fig. 72.—Cyclonic wind velocity, winter, northern United States.

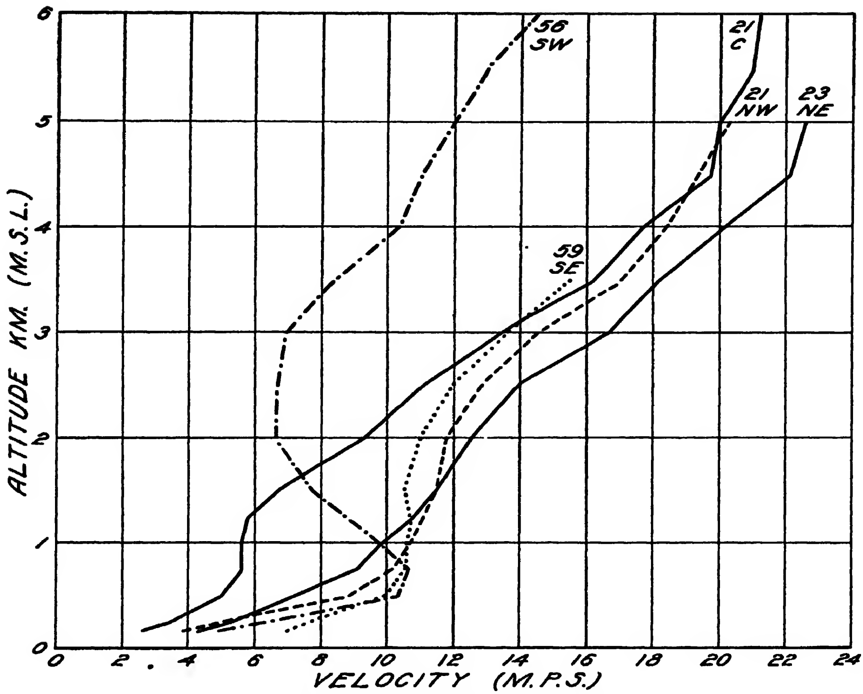

Fig. 73.—Anticyclonic wind velocity, winter, southern United States.

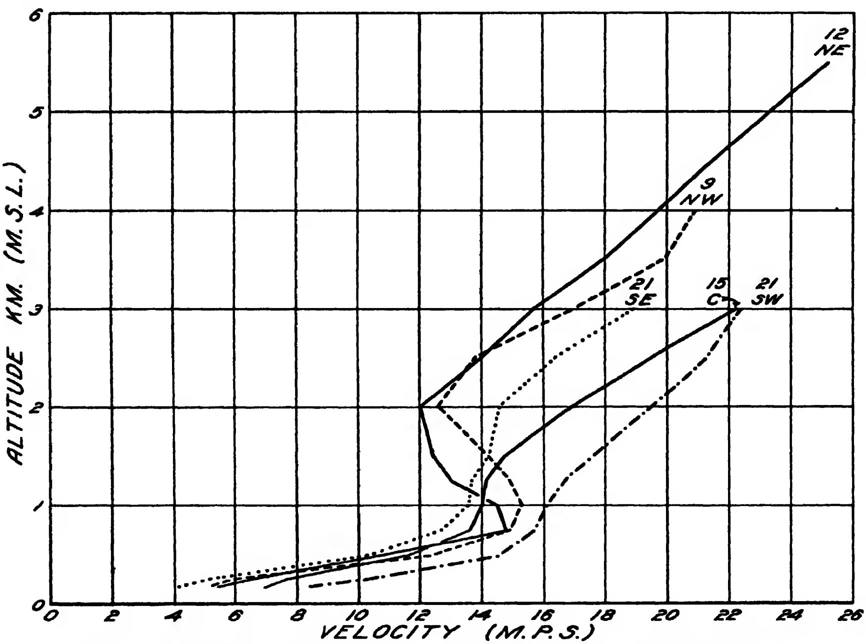

Fig. 74.—Cyclonic wind velocity, winter, southern United States.

5) The difference between the mean velocities at the northern stations and those at the southern stations is considerably less during winter than during summer.

6) The difference between the mean velocities for the various quadrants is considerably greater in the higher levels than at the surface and lower levels, where these differences are exceedingly small.

7) There is very little difference, in the average wind velocities in winter HIGHS between the northern and southern sections of the United States. (Figures 71 and 73.)

8) Most conspicuous in the mean velocities of the various quadrants are the comparatively light winds in the higher levels of the SW quadrant and in the lower levels of the central region.

9) The strongest winds in the higher levels in winter HIGHS are found in the SE quadrant in northern United States and in the NE quadrant in southern United States.

10) Figures 72 and 74 show a number of marked differences in the mean wind velocities in the higher levels of winter LOWS between northern and southern United States. The rate of increase in velocity with elevation is less pronounced in the North than in the South.

11) Summer conditions, especially in the southern part of the United States, present a sharp contrast to winter conditions, in that the rates of increase in velocity with elevation are very small during summer. An exception is found, however, in the case of summer HIGHS in the North, where the velocities become quite strong in the upper levels in the NE and SE quadrants.

Characteristics of the free-air conditions in HIGHS and LOWS.—In a summary of aerological observations made in well-pronounced HIGHS and LOWS based on observations in the United States east of the Rocky Mountains, the following results were found: [33]

1) The characteristics distinguishing HIGHS from LOWS are most pronounced in their *lower levels*.

2) The front sector * of HIGHS should not be considered as synonymous with the rear sector of LOWS nor should the rear sector of HIGHS be so considered with respect to the front sector of LOWS, since the mean values of the various elements in these respective regions show distinct differences which, however, decrease with elevation until they become very small above 5 km.

* The term "front sector" (NE and SE quadrants) is here used to designate the region of HIGHS where the general wind direction in the lower levels is northerly and in LOWS, southerly. The "rear sector" (NW and SW quadrants) refers to the region of HIGHS where this direction is southerly and in LOWS, northerly. Quadrants are numbered counterclockwise, *i. e.*, NE = I, NW = II, etc.

3) The rear sector of HIGHS contains a much greater change in the mean wind direction with elevation than their front sector, whereas in LOWS this change is greatest in the northeast and northwest quadrants and least in the southwest and southeast quadrants.

4) A common mean wind direction for all quadrants of HIGHS is reached at a greater height in summer than in winter and at successively higher altitudes from northern to southern stations.

5) The average temperature lapse rates in both the front and rear sectors of HIGHS and LOWS are greater in summer than in winter, the seasonal differences being much greater for HIGHS than for LOWS.

6) The average lapse rate is greater in LOWS than in the adjacent sector of HIGHS for the same season, the differences being greater in winter than in summer.

7) In summer, LOWS average warmer than the adjacent sectors of HIGHS, the differences becoming progressively smaller from northern to southern stations and from lower to higher altitudes.

8) In winter, LOWS average warmer than HIGHS at the northern stations but generally colder at the southern and eastern stations at their upper levels, particularly in their rear sector as compared with the front sector of HIGHS.

9) The front sector of HIGHS in winter averages colder than the rear sector to at least 5 km., the differences decreasing from northern to southern stations. For the same season, the front sector of LOWS averages warmer than the rear, to at least 4 km. However, in this case the differences increase from northern to southern stations.

10) The differences between the mean relative humidities for adjacent sectors of HIGHS and LOWS at northern stations are small, but slightly lower humidities are indicated, in general, for LOWS than for HIGHS in both seasons. The opposite relationship is found, however, at the southern and eastern stations in winter, where the relative humidities average higher in LOWS than in HIGHS.

11) The relative humidity in HIGHS and in LOWS at the northern stations averages a little higher in winter than in summer, whereas the opposite relationship occurs in the case of HIGHS at the southern stations.

12) The relative humidity in the upper levels of HIGHS averages highest in the second quadrant and lowest in the central region. In general, at all levels, it averages higher in the fourth quadrant of HIGHS than in the third at the northern stations but the opposite relationship occurs at the southern stations.

13) The mean vapor pressures in both seasons are appreciably greater in LOWS than at the same levels of the adjacent sector of HIGHS.

14) The mean vapor pressures in HIGHS and in LOWS are greater at the southern stations than at the northern for the same seasons and levels. This latitudinal variation is appreciably less for LOWS than for HIGHS in both seasons.

15) The mean vapor pressures of the front sector of HIGHS average lower than those of their rear sector both at the northern and southern stations in both seasons, whereas in LOWS the mean values are, in general, lowest in the rear sector. These differences, in both pressure systems, diminish with altitude.

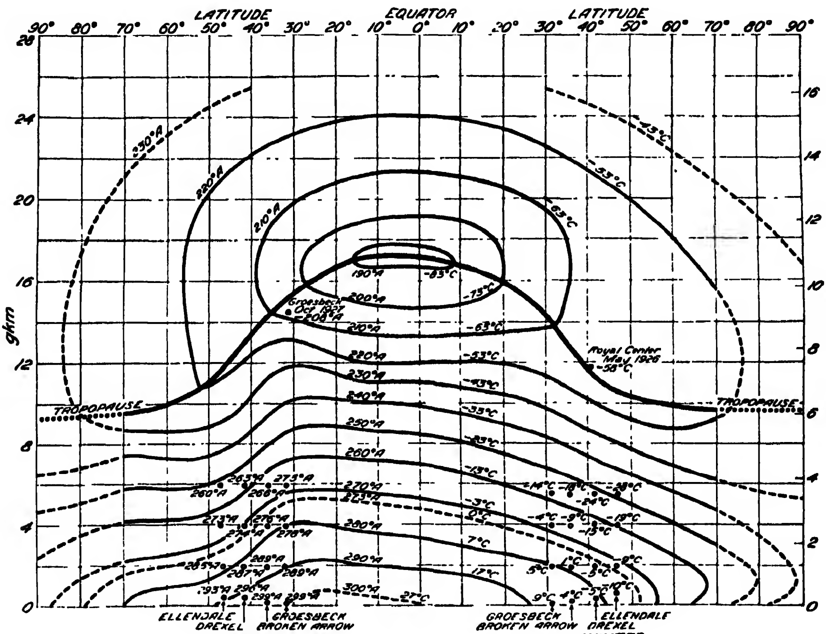

FIG. 75.—Distribution of temperature up to 25 kilometers over the northern hemisphere during summer and winter.

Distribution of temperature up to 25 km. over the Northern Hemisphere.—Figure 75 has been taken from an article published by K. R. Ramanathan, Meteorological Department, Poona, India, in Nature (London) June 1, 1929. There have been added to this figure, for comparison purposes, the average height and temperature of the tropopause as determined from a sounding balloon series made at Royal Center, Indiana, in May 1926 and at Groesbeck, Texas, in October 1927. There are also indicated the average temperatures at various heights as determined for summer and winter from kite observations made in the United States. Although all of the latter values do not coincide with the smoothed iso-

therms of Ramanathan's chart, the general agreement is good, and the differences found are doubtless real and are due to the greater temperature extremes occurring in continental United States as compared with Europe.

Regarding this Chart, Ramanathan states as follows:

(1) The stratosphere is not isothermal over any particular place, but above a certain level there is a tendency for the temperature to increase with height.

(2) The coldest air over the earth, of temperature about 185° A. (− 88° C.), lies at a height of some 17 geodynamic kilometers* (17.4 geometric kilometers) over the Equator in the form of a flat ring surrounded by rings of warmer air.

(3) The surface of the tropopause has a relatively steep slope toward the pole between latitudes 30° and 50° in summer and between 25° and 45° in winter.

(4) The ring of lowest temperature at the tropopause is displaced towards the summer hemisphere.

(5) There is a ridge of high temperature in the tropopause between latitudes 20° and 40° N. in summer, corresponding to the ridge of high pressure at 8 km. over those latitudes. (See Sir Napier Shaw's chart of 8-km. isobars in July, Manual of Meteorology, Vol. 2, p. 262.)

DIURNAL VARIATION OF TEMPERATURE

Surface.—As the sun rises above the horizon the amount of radiant energy received at the earth's surface becomes greater than the amount radiated out from the earth. The temperature rises as long as this excess continues, or, as a rule, until sometime in the afternoon. The temperature then begins to fall, since the terrestrial radiation exceeds the radiation received, and continues to fall until shortly after sunrise of the following day.

Hourly means over a long period of time show a very regular variation, while as a result of the advance of atmospheric disturbances, individual days often show an irregular variation.

At the equator the amount of heat received from the sun varies but little from season to season; furthermore, day and night are always

* With reference to the "geodynamic kilometer" scale used in Figure 75, the following is quoted from Manual of Meteorology, 2: XXI, by Sir Napier Shaw.

"There is a good deal of laxity about the use of the word height, of the same kind as that of the aeronauts who graduate a pressure-instrument to read what they call height. For example, V. Bjerknes and others would express the height of a point in the atmosphere by the geopotential at the point, calling the quantity expressed, the dynamic height. We reproduce from the *Avant propos* of the *Comptes rendus des jours internationaux,* 1923: 'The relation of the geopotential at any position to the geometric height of that position h and the gravitational acceleration g is $\Gamma = \int g dh$. The value is governed accordingly by the local value of gravity depending on the attraction of gravitation and the rotation of the earth; but not to any appreciable extent upon the condition of the atmosphere at the time of observation.'"

equal in length; so that the amplitude of the diurnal variation remains practically constant. In summer, in temperate latitudes, the heating of the earth is considerable during the day, and since the amount of heat radiated by a body is directly proportional to the fourth power of its absolute temperature, the cooling at night is rapid, resulting in a comparatively large diurnal amplitude. In winter, on the other hand, the amount of solar radiation received is much less, consequently the warming of the surface is less and the cooling at night is slower. A 24-hour period in temperature obviously does not exist at the poles.

Besides latitude and season there are a number of factors that locally affect the amplitude of the diurnal temperature, such as clouds, topography, and vegetation.

When the sky is overcast, the amount of heat received during the day at the earth's surface is greatly diminished. At night, on the other hand, an overcast sky absorbs a large amount of the terrestrial radiation and re-radiates it back to the surface. The effect, then, of an overcast sky is to reduce the amplitude.

A station situated in a valley will have a larger diurnal variation than one at the same level and latitude situated on a plain. During the day the reflection of the sun's rays from the sides of the mountain warms up the air more than on a plain, while at night the mountain sides cool by radiation, and the air cooled by contact with them drains downward to the valleys, causing a lower temperature than on a plain.

A station situated far inland will have a greater diurnal variation than one on an island in the ocean. Part of the radiation received at the surface of the ocean is used to raise the temperature of the water, and a part is used to evaporate the water without change in temperature. Furthermore, the specific heat of water is considerably greater than that of the soil, so that the same amount of heat will cause a greater rise in temperature at a continental than at an ocean station. Because of its higher temperature, lower specific heat, and less water-vapor content, the loss of temperature at night will be more rapid at a continental station than at an island station.

A station in a desert region will have a greater diurnal amplitude in temperature than one situated in a region where there is a great amount of vegetation. In the former, the radiant energy goes to heat the barren surface. In the latter case, much of the energy is absorbed by the vegetation or used to evaporate moisture. Since the water-vapor content is greater in a region where vegetation is rich, more terrestrial radiation will be re-radiated back to the surface at night; and since the temperature is lower, the rate of fall of temperature will be less than at a desert station.

Aloft.—Since solar radiation passes through the atmosphere without heating it appreciably, the absorption of solar energy is concentrated at the earth's surface. The heated surface warms the air in contact with it, partly by conduction, and partly by absorption of the long wave-length radiation emitted by the earth's surface. When the heating of the lower layers is sufficient to produce a lapse rate greater than the adiabatic, convection takes place and the rising air carries its surplus heat with it, resulting in an increase in temperature at higher levels. Obviously, some heat is conveyed upward by terrestrial radiation, but the major portion of the transfer takes place by convection. At night, if the air is very dry, cooling is most rapid at the earth's surface, because of its higher temperature and also because it is a better radiator than the atmosphere. If the air is very moist a large percentage of the terrestrial radiation is absorbed and re-radiated back to the surface, thus tending to equalize the temperature between day and night.

Obviously, the effect of the foregoing phenomena is to cause a decrease in the diurnal variation of temperature with increasing height.

Table 9 shows the diurnal variation over Drexel, Nebraska, the figures representing departures from the mean temperature for the season and level in question. The data were obtained from a total of 158 kite flights which were made over a number of 24-hour periods, averaging eight or nine flights per period.

It will be noted that the amplitude decreases very rapidly in the lower levels with increase in altitude becoming almost negligible at about 1,500 meters.

In summer and autumn there is a retardation in the time of the occurrence of the maximum temperature with increase in altitude at all levels studied, but in spring the lag occurs only to an altitude of 2,000 meters and in winter to 1,000 meters. At higher levels the maximum temperature in the latter seasons occurs at an earlier hour than at the immediately preceding levels.

TABLE 9

Diurnal Variation of Temperature at Drexel, Nebraska

Surface (396 M.)

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	3.6	4.9	5.8	6.5	6.3	5.8	4.6	2.5	1.0	−0.1	−1.0	−1.9	−2.6	−3.3	−3.9	−4.6	−4.9	−5.3	−5.5	−4.9	−3.5	−1.5	0.7	2.3
Summer	3.2	4.3	5.0	5.3	5.2	4.9	4.0	2.3	0.5	−0.7	−1.4	−2.0	−2.6	−3.1	−3.6	−4.2	−4.7	−5.0	−4.7	−3.6	−2.3	−0.6	0.8	2.1
Autumn	3.6	5.1	6.0	6.5	6.2	4.8	3.1	1.6	0.4	−0.3	−1.0	−1.6	−2.2	−2.0	−3.5	−4.0	−4.4	−5.0	−5.2	−4.9	−3.6	−1.7	0.2	1.9
Winter	1.7	3.1	4.2	5.0	4.9	3.9	2.6	1.5	0.8	0.2	−0.4	−0.8	−1.1	−1.6	−1.9	−2.3	−2.7	−3.0	−3.3	−3.6	−3.6	−2.5	−1.2	0.3
Annual	3.0	4.8	5.3	5.8	5.6	4.8	3.5	2.0	0.8	−0.2	−0.9	−1.5	−2.0	−2.5	−3.0	−3.6	−4.1	−4.4	−4.5	−4.2	−3.2	−1.7	0.0	1.6

750 M.

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	−0.2	0.9	1.8	2.4	2.7	2.7	2.3	2.0	1.7	1.1	0.9	0.6	0.2	−0.1	−0.4	−0.7	−1.3	−1.8	−2.2	−2.6	−2.6	−2.6	−2.0	−1.1
Summer	−0.1	0.7	1.5	2.1	2.4	2.1	2.1	1.8	1.5	1.3	0.9	0.4	−0.1	−0.6	−0.9	−1.2	−1.5	−1.8	−2.0	−2.3	−2.4	−2.3	−1.6	−1.0
Autumn	−1.0	0.5	1.3	2.0	2.0	2.0	1.9	1.6	1.4	1.3	1.1	0.7	0.2	−0.1	−0.4	−0.8	−1.0	−1.3	−1.5	−1.7	−1.8	−1.9	−1.8	−1.6
Winter	−0.5	0.0	0.6	1.0	1.2	1.1	1.1	0.9	0.9	0.8	0.7	0.7	0.5	0.4	0.3	0.0	−0.4	−0.8	−1.1	−1.4	−1.6	−1.5	−1.1	−0.8
Annual	−0.6	0.4	1.2	1.7	1.9	1.8	1.8	1.5	1.2	1.1	0.8	0.6	0.2	−0.1	−0.4	−0.7	−1.0	−1.5	−1.7	−2.0	−2.1	−2.0	−1.7	−1.2

1,000 M.

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	−0.6	0.3	0.9	1.4	1.6	1.7	1.6	1.4	1.2	0.7	0.6	0.5	0.4	0.3	−0.1	−0.3	−0.7	−1.1	−1.3	−1.6	−1.6	−1.6	−1.6	−1.3
Summer	−0.7	−0.2	0.5	1.1	1.2	1.3	1.4	1.3	1.0	0.8	0.7	0.4	0.0	−0.2	−0.4	−0.5	−0.7	−0.8	−0.9	−1.1	−1.3	−1.4	−1.4	−1.2
Autumn	−0.8	−0.3	0.1	0.5	0.7	0.8	0.8	0.6	0.5	0.6	0.7	0.6	0.3	0.2	0.1	0.0	−0.2	−0.5	−0.7	−0.7	−0.9	−1.0	−1.0	−1.0
Winter	−0.2	0.1	0.4	0.4	0.4	0.6	0.5	0.2	0.3	0.3	0.3	0.3	0.3	0.4	0.3	0.1	0.0	−0.2	−0.6	−0.7	−0.9	−0.8	−0.4	−0.3
Annual	−0.6	−0.1	0.5	0.7	0.7	1.0	1.0	0.8	0.6	0.6	0.5	0.5	0.3	0.1	0.0	−0.2	−0.4	−0.7	−0.9	−1.1	−1.1	−1.1	−1.1	−0.9

1,500 M.

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	−0.5	−0.3	0.0	0.2	0.4	0.4	0.3	0.1	0.2	0.3	0.4	0.5	0.6	0.6	0.5	0.2	0.0	−0.4	−0.5	−0.6	−0.6	−0.7	−0.7	−0.8
Summer	−0.2	−0.1	0.1	0.3	0.3	0.4	0.6	0.5	0.4	0.3	0.2	0.1	0.1	0.0	−0.2	−0.2	−0.2	−0.2	−0.2	−0.3	−0.3	−0.2	−0.1	−0.1
Autumn	−0.3	−0.1	0.1	0.3	0.6	0.6	0.5	0.2	0.1	0.2	0.3	0.2	0.1	0.1	−0.1	−0.1	−0.1	−0.2	−0.3	−0.4	−0.3	−0.2	−0.3	−0.3
Winter	−0.1	0.1	0.2	0.4	0.5	0.5	0.4	0.2	0.1	0.1	0.2	0.1	−0.1	−0.1	−0.1	−0.1	−0.2	−0.3	−0.3	−0.3	−0.3	−0.2	−0.1	−0.1
Annual	−0.3	−0.1	0.1	0.2	0.4	0.4	0.4	0.3	0.1	0.2	0.2	0.2	0.2	0.1	0.0	−0.1	−0.2	−0.4	−0.4	−0.4	−0.4	−0.3	−0.3	−0.3

TABLE 9—*Continued*

2,000 M.

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	−0.3	−0.2	0.0	0.0	0.2	0.3	0.3	0.1	0.1	0.1	0.3	0.4	0.4	0.2	−0.1	−0.1	−0.2	−0.4	−0.5	−0.4	−0.3	−0.3	−0.4	−0.3
Summer	−0.1	0.0	0.1	0.1	0.2	0.2	0.1	0.1	0.2	0.0	0.0	0.0	0.0	0.0	−0.2	−0.3	−0.4	−0.2	−0.2	−0.3	−0.1	0.0	0.0	−0.1
Autumn	−0.1	0.0	0.0	0.2	0.3	0.3	0.2	0.1	0.2	0.2	0.3	0.2	0.2	0.1	0.0	0.0	0.0	0.0	−0.2	−0.3	−0.3	−0.2	−0.1	−0.1
Winter	0.3	0.5	0.7	0.7	0.7	0.5	0.4	0.1	0.0	−0.1	−0.1	−0.2	−0.3	−0.4	−0.5	−0.5	−0.5	−0.4	−0.3	−0.2	−0.1	−0.1	0.0	0.1
Annual	0.0	0.0	0.2	0.3	0.3	0.3	0.3	0.1	0.1	0.1	0.0	0.1	0.0	−0.1	−0.2	−0.2	−0.3	−0.3	−0.3	−0.3	−0.2	−0.1	−0.1	−0.1

2,500 M.

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	−0.2	0.1	0.1	0.2	0.5	0.4	0.2	0.1	0.0	0.0	0.0	0.1	0.1	−0.1	−0.1	−0.2	−0.1	−0.3	−0.3	−0.3	−0.4	−0.3	−0.3	−0.3
Summer	0.0	0.1	0.0	0.1	0.1	0.1	0.2	0.4	0.4	0.2	0.0	−0.1	−0.1	−0.1	−0.2	−0.2	−0.4	−0.5	−0.5	−0.4	−0.2	−0.1	0.1	0.1
Autumn	−0.2	0.0	0.1	0.1	0.2	0.2	0.3	0.3	0.4	0.3	0.3	0.1	0.1	0.1	0.0	0.0	−0.1	−0.3	−0.4	−0.4	−0.5	−0.3	−0.2	−0.2
Winter	0.3	0.7	0.8	0.8	0.7	0.4	0.3	0.2	−0.1	−0.1	−0.2	−0.4	−0.4	−0.5	−0.5	−0.4	−0.4	−0.5	−0.4	−0.2	−0.2	0.0	0.2	0.3
Annual	0.0	0.2	0.2	0.3	0.4	0.3	0.3	0.2	0.1	0.1	0.0	−0.1	−0.1	−0.2	−0.2	−0.2	−0.3	−0.4	−0.5	−0.3	−0.3	−0.2	−0.1	−0.1

3,000 M.

Hour	12	1p	2	3	4	5	6	7	8	9	10	11	12	1a	2	3	4	5	6	7	8	9	10	11
Spring	0.3	0.7	0.8	0.8	0.7	0.6	0.4	0.2	0.1	0.0	−0.1	−0.1	−0.2	−0.3	−0.4	−0.5	−0.4	−0.4	−0.3	−0.2	−0.2	−0.2	−0.2	0.1
Summer	−0.2	−0.1	0.0	0.1	0.2	0.2	0.3	0.3	0.2	0.3	0.1	0.0	0.0	0.0	0.0	0.1	0.2	0.0	−0.1	−0.2	−0.1	−0.2	−0.2	−0.2
Autumn	−0.1	0.2	0.2	0.2	0.1	0.1	0.1	0.2	0.4	0.4	0.4	0.3	0.1	0.1	0.0	−0.1	−0.3	−0.4	−0.5	−0.6	−0.6	−0.4	−0.1	−0.1
Winter	0.6	0.5	0.5	0.5	0.4	0.3	0.2	0.1	0.0	0.0	−0.1	−0.4	−0.5	−0.6	−0.7	−0.6	−0.5	−0.4	−0.3	−0.2	−0.1	0.2	0.5	0.6
Annual	0.2	0.3	0.4	0.3	0.3	0.2	0.2	0.2	0.2	0.1	0.1	−0.1	−0.1	−0.2	−0.3	−0.3	−0.3	−0.3	−0.4	−0.3	−0.3	−0.1	0.0	0.1

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CHAPTER V

DYNAMIC METEOROLOGY

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FOREWORD

The purpose of the present paper is to outline briefly, and in as far as possible non-mathematically, the outstanding features of modern dynamic meteorology. Lack of space permits of consideration only of the leading schools of thought. The aim has been to select in each particular field the best work which has appeared in that field and to develop a logical and consistent treatment of the results. Especial attention is paid to the work of Bjerknes and the Norwegian school, of Exner and the Austrian school, and of certain of the German meteorologists, for many of the most important recent contributions to the science of meteorology have come from these sources, and have made their way but slowly into English meteorological literature.

The use of mathematics has been kept at a minimum. Consequently, only the most fundamental relationships are developed mathematically, and those only in the simplest and most obvious way. Owing to lack of space applications of the general principles discussed to specific atmospheric phenomena are necessarily few. Only the general or primary circulation, and the secondary, or cyclonic and anticyclonic circulations are treated in any detail. The numerous local phenomena, such as the thunderstorm, tornado, föhn, drainage winds, and many others, have received no attention.

I. THE GAS LAWS AS APPLICABLE TO THE ATMOSPHERE

Dynamic meteorology is concerned with the quantitative treatment of the forces determining atmospheric changes. It is a subject of great practical value as well as theoretical interest. Its laws, when perfected by means of a wider range of observations than is now available, especially of upper air phenomena, will make possible scientific, as distinct from empirical, weather forecasting. The importance of the subject from this point of view, especially in its relation to aviation, can hardly be overestimated.

The atmosphere, the scene of all meteorological phenomena, is a mixture of gases, subject to the laws applying to gases in general. The application of these laws to atmospheric phenomena is complicated, however,

by the presence of the variable constituent water vapor. Apart from this element the composition of the atmosphere may be considered as constant up to at least 20 kilometers, a limit beyond which the application of these laws has little if any practical meteorological significance. This constant mixture of gases is customarily referred to simply as dry air. The temperature, pressure and density relationships holding for dry air will be discussed first, and then the necessary modifications will be made to allow for the presence of water vapor and finally for the process of condensation.

The equation of state for dry air.—The equation of state for any gas or constant mixture of gases expresses the relationship existing between the temperature, pressure and density of the gas.

For dry air the combination of Boyle's and Charles' laws gives the relationship $pv=p_0v_0(1+\alpha t)$ where p and v are pressure and volume (the subscript $_0$ referring to initial conditions), $\alpha=\frac{1}{273}$, and t is temperature in degrees centigrade. Introducing the absolute temperature T, this may be written $pv=p_0v_0\frac{T}{273}$, or if the standard conditions of 760 mm. pressure and 0° C. temperature are used as reference, $pv=RT$, where $R=\frac{p_0v_0}{T_0}$. If v_0 refers to the specific volume (volume of unit mass) of the gas in question under standard conditions, R becomes a characteristic constant for any gas or constant mixture of gases, such as dry air, and is known as the gas constant. Since R varies only with v_0, which by Avogadro's law varies inversely as the molecular weight of the gas, $R=\frac{R'}{M}$, where R' is the universal gas constant, and M the molecular weight of the gas or mixture of gases in question. The relationship $pv=RT$, or $p=\rho RT$ is called the equation of state of a gas.

Adiabatic changes in dry air.—Poisson's equation expresses the relationship between the temperature and pressure of a gas or constant mixture of gases in the absence of all transfer of heat to or from the gas.

The first law of thermodynamics states that if a quantity of heat is supplied to unit mass of dry air it is used in part to raise the temperature of the air, and in part in performing the work of expansion against the external pressure incident to the heating of the air mass. Expressed quantitatively $dQ=C_vdT+Apdv$, where dQ is the amount of added heat, C_v the specific heat of air at constant volume, and A the heat equivalent of work. The volume differential dv may be eliminated between this equation and the equation of state, which gives $pdv+vdp=RdT$ and, upon substitution, $dQ=(C_v+AR)dT-Avdp$. If the pressure is con-

stant, or $dp=0$, it follows that $dQ=C_p dT$, where C_p is the specific heat at constant pressure. Hence $C_v+AR=C_p$, and $dQ= C_p dT-Avdp=C_p dT -ART\frac{dp}{p}$ and, dividing by T, $\frac{dQ}{T}=C_p\frac{dT}{T}-AR\frac{dp}{p}$.

The expression on the right is the perfect differential of the quantity $S=C_p \log T-AR \log p$, which in thermodynamics is called entropy. Hence $\int\frac{dQ}{T}=S+\text{constant}$. The condition that there be no addition or subtraction of heat to or from the unit air mass, or $dQ=0$, gives the *adiabatic* relationship between the temperature and pressure of any portion of dry air subject to changing conditions. If the change is assumed to be adiabatic, then $C_p\frac{dT}{T}=AR\frac{dp}{p}$, which gives upon integration between initial conditions p_0 and T_0, and end conditions p and T, $C_p \log\frac{T}{T_0}=AR \log\frac{p}{p_0}$, or $\frac{T}{T_0}=\left(\frac{p}{p_0}\right)^{\frac{AR}{Cp}}=\left(\frac{p}{p_0}\right)^{0.2884}$ which is known as Poisson's equation. If instead of eliminating the specific volume v between the equation of state and the heat equation, the pressure p had been eliminated, a similar relationship would have been found: $\frac{T}{T_0}=\left(\frac{\rho}{\rho_0}\right)^{\frac{AR}{Cv}}$, where $\rho=\frac{1}{v}$ is the density.

From $\left(\frac{\rho}{\rho_0}\right)^{\frac{AR}{Cv}}=\left(\frac{p}{p_0}\right)^{\frac{AR}{Cp}}$, it follows that $\left(\frac{p}{p_0}\right)=\left(\frac{\rho}{\rho_0}\right)^{\frac{Cp}{Cv}}$. Since atmospheric movements take place as a rule, in the absence of condensation, with comparatively slight addition or removal of heat, the above relationships find very general application. As will be shown presently, the presence of water vapor does not materially affect these dry air relationships until condensation begins.

In order to compare the actual heat condition of air masses under different pressures and temperatures, it is customary to use the so-called *potential temperature* θ, which is the temperature which an air mass will have if reduced adiabatically to a standard pressure of 1,000 millibars or one of 760 mm. In terms of the initial temperature and pressure it is to be expressed $\theta=T\left(\frac{760}{p}\right)^{\frac{AR}{Cp}}$, the constant exponent $\frac{AR}{C_p}$, having approximately the numerical value 0.2884.

This equation defining θ may also be written $\log\theta=\log T-\frac{AR}{C_p}\log p+\frac{AR}{C_p}\log 760$, or $C_p\log\theta=C_p\log T-AR\log p+\text{constant}=S+\text{con-}$

stant, whence $S=C_p \log \theta+\text{constant}$, where S is entropy as defined above. This equation expresses the important fact that the entropy of an air mass changes proportionally to the potential temperature, both remaining constant for adiabatic changes.

The equation of state for moist air.—It is necessary next to consider the effect of water vapor on the temperature, pressure, and density relationships in the atmosphere. In the atmosphere there is always some water vapor present, though the amount is exceedingly variable. In the vapor state, as long as no condensation occurs, water vapor behaves just as any of the other gases whose partial pressures constitute the total atmospheric pressure p. The amount of vapor present is usually expressed in terms of the partial pressure e, in mm. of mercury. Experiment has shown that for each temperature $t°$ C. there is a certain maximum vapor pressure e_m at which condensation begins. This maximum or *saturation pressure* is given approximately by the formula of Magnus:

$$e_m=4.525\times 10^{\frac{7.4475t}{234.67+t}} \text{ mm. mercury,}$$

though for exact values of e_m it is necessary to go to one of the numerous tables prepared for the quantity, such as the Smithsonian tables.

If e remains $<e_m$, and ρ_w is the water vapor density due to the vapor pressure e, the equation of state for the vapor is $e=\rho_w R_w T=1.605\rho_w RT$, since the molecular weight of dry air is 1.605 times that of water vapor. The water vapor density $\rho_w=\dfrac{e}{R_wT}$ is defined as the *absolute humidity*. The ratio of the mass of water vapor to the total mass of damp air per unit volume, or $q=\dfrac{\rho_w}{\rho}$, is defined as the *specific humidity*. The equation of state for moist air is similar to that for dry air, the only difference being that the gas constant R for dry air must be increased by an amount $q(1.605-1.000)$ to allow for the different molecular weight of the new constituent water vapor. Hence the equation for unsaturated moist air may be written $p=\rho R(1+0.605q)T$, where p, ρ, and T refer to the moist air, and R is the gas constant for dry air. It may also be written $p=\rho RT'$, where $T'=T(1+0.605q)$, and is defined as the *virtual temperature*. The advantage of the use of virtual temperature is that the equation of state then retains the same form for moist as for dry air, which is of practical value in certain computations especially in ballistics. Since $T'>T$, it follows that moist air will be lighter than dry air at the same temperature and pressure in proportion to its water vapor content, a fact which is evident directly from the smaller molecular weight of water vapor.

From $\frac{e}{p}=\frac{1.605\rho_w RT}{(1+0.605q)\rho RT}$, remembering that $\frac{\rho_w}{\rho}=q$, it follows that $q=\frac{e}{1.605p-0.605e}$, or approximately, since e is small compared with p, $q=\frac{5}{8}\frac{e}{p}$. This makes q a readily determined and quite conservative characteristic of any unsaturated air mass.

The ratio $f=\frac{e}{e_m}$ expressed as a percentage is defined as the *relative humidity.* The relative humidity f together with the specific humidity q and absolute humidity ρ_w are the quantities commonly used as measures of the moisture content of the atmosphere.

Adiabatic changes in moist air.—If a portion of unsaturated moist air is subjected to adiabatic changes of pressure and temperature, the relation between temperature and pressure is less simple than that for dry air expressed by Poisson's equation. If the pressure is continuously reduced, the changes taking place in the moist air mass fall naturally into four stages, which are best considered separately, as was first done by Hertz, more thoroughly by Neuhoff, and recently by Humphreys[1] and Fjeldstad.[2]

(a) *The dry stage.*—With the continued reduction of p, and therefore of T, the temperature of the moist air mass must eventually reach that temperature t, its saturation temperature, for which e_m as given by Magnus' formula, is equal to the e prevailing in the moist air. In the dry stage, which ends with the attainment of the saturation temperature, it was found that $pv=R'T$, where $R'=R(1+0.605q)$. Likewise $dQ=C'_p dT+Apdv$, where C'_p applies to the moist air. Since $C'_p=(1-q)C_p+qC_{p_w}$, and $\frac{C_{p_w}}{C_p}=1.596$, it follows that $C'_p=(1+0.596q)C_p$. The adiabatic relationship between T and p may be derived just as for dry air, unit mass of moist air containing the mass q of water vapor. It follows that

$$\frac{T}{T_0}=\left(\frac{p}{p_0}\right)^{\frac{AR'}{C'_p}}=\left(\frac{p}{p_0}\right)^{\frac{AR(1+0.605q)}{C_p(1+0.596q)}},$$

and since the ratio q even in the tropics rarely exceeds $\frac{1}{40}$, the exponent may be written $\frac{AR}{C_p}(1+0.009q)$. The factor $(1+0.009q)$ is of the same order of magnitude as the probable error in the best determined value of the constant $\frac{AR}{C_p}$, hence it may be neglected entirely, and even for the most exact reckoning Poisson's equation for dry air holds for the unsaturated stage in moist air.

(b) *The rain stage.*—From the point at which condensation begins in the expanding and cooling moist air mass until the temperature reaches the freezing point and the condensed water begins to freeze, the system is in the rain stage. If it happens that the vapor pressure is sufficiently small the saturation temperature will be below freezing, and the system passes directly from the dry stage to the snow stage (d). As soon as the rain stage is reached, further cooling by adiabatic expansion is accompanied by condensation, which in turn reduces the rate of cooling in proportion to the rate of condensation. Since there is to be no gain or loss of heat from the outside, it must be assumed that all the condensed water is retained within the system. There must then be equality between the heat used in expansion on one side, and the heat supplied by condensation, and by the cooling of the air, the saturated water vapor, and the condensed water on the other hand. To a fairly close approximation * the specific heats of the water vapor and condensed water may be neglected, and the equation corresponding to the Poisson equation for dry air may be written $C_p \log T - AR \log p + 0.623 \frac{re_m}{pT} = \text{constant}$, where r is the heat of condensation, and e_m refers to the temperature T. Evidently the lower the temperature the smaller is e_m, or the less the heat supplied by condensation, and the more nearly does the *saturation adiabatic* relationship between T and p approach the dry.

(c) *The hail stage.*—When the adiabatically expanding moist air reaches the freezing point, the condensed water begins to freeze. The heat of fusion set free keeps the temperature constant until the liquid water is entirely frozen. At the same time, since e_m remains constant and the volume is expanding, a certain amount of the liquid water must be evaporated to maintain the vapor pressure. Hence there must be equality between the heat of fusion set free and the heat consumed in evaporation and expansion. Integration between the limits p_0 and p_1 which respectively indicate the initial and final pressure of the hail stage gives the following relationship:

$$AR \log \frac{p_0}{p_1}\left(\frac{r+K}{p_1} - \frac{r}{p_0}\right)\frac{0.623e_m}{T} = \frac{Kq}{T} = 0,$$

where K is the heat of fusion, and $r+K$, theoretically, the heat of sublimation. p_0 and T are determinable from the initial conditions of the adiabatically expanding air mass, therefore p_1, the only unknown in the above equation, may be directly calculated.

* For a rigid development of the complete equations for these four stages see Fjelstad's paper[2] on adiabatic changes of condition for moist air.

(d) *The snow stage.*—After the freezing of all fluid water, further adiabatic expansion leads to direct sublimation and further temperature fall. The same heat balance exists as in the rain stage, the only difference being that now instead of the specific heat of fluid water the specific heat of ice is to be considered, and instead of the heat of condensation, there is set free the heat of sublimation. The approximation formula corresponding to that of the rain stage is the following:

$$C_p \log T - AR \log p + \frac{0.623(r+K)e_m}{pT} = \text{constant.}$$

At very low temperatures the saturation pressure e_m approaches zero, and the formula becomes that of Poisson's equation. Fjeldstad[3] has found reason to think that actually the heat of sublimation is not equal to $r+K$, where r is assumed to increase with decreasing temperature at sub-zero temperatures as it does above freezing, but that it has the constant value 680, the zero value. His thermodynamic diagram is computed on the basis of this constant heat of sublimation.

Pseudoadiabatic changes.—It was pointed out that if changes of condition in moist air are to be truly adiabatic, all the products of condensation (rain, hail, or snow) must be retained within the system. If it is assumed that they are completely removed as soon as formed, which probably corresponds more closely to what occurs in the atmosphere, the changes are called *pseudoadiabatic.* In the stages discussed above, this assumption has the effect of dropping out the terms involving the specific heat of water and ice in stages *b* and *d* (which have already been dropped out of the shortened equations given there) and of eliminating the hail stage completely by preventing the accumulation of liquid water capable of freezing. Adiabatic changes are reversible, but pseudoadiabatic are evidently irreversible, for once the condensed water is eliminated, there can be no removal of heat by evaporation such as would be necessary to reverse the initial supply of heat by condensation.

Condensation can begin in the atmosphere only in the presence of dust particles or particles of hygroscopic matter which can act as nuclei to hold the water molecules together. The effect of surface tension acting over exceedingly small droplets renders them unstable without some such nucleus. Recent extensive investigations by H. Köhler[4] of the nuclei of condensation and cloud droplets have led him to conclude that the condensation processes in the atmosphere actually depart considerably from the four ideal stages outlined above. His idea of the condensation processes is briefly as follows:

(a) The condensation nuclei are primarily particles of sea salt or other moderately hygroscopic substances which are ever present in the

atmosphere and which hold attached increasing amounts of condensed water as the relative humidity approaches 100 per cent.

(b) These droplets, or cloud particles in general, do not freeze as the adiabatically expanding air mass falls below freezing, but remain to quite low temperatures as supercooled water droplets, consequently there is no definitely bounded hail stage, and condensation rather than sublimation continues far below freezing.

(c) The temperature at which the individual droplets freeze depends upon the size to which they grow and the mechanical disturbances (atmospheric turbulence and jostling together) to which they are subjected. Direct sublimation takes place only as the supercooled droplets congeal, after which they serve as nuclei for the sublimation.

There are many direct and indirect observations confirming this point of view. That clouds frequently consist of supercooled droplets even at temperatures far below freezing is indicated by many direct mountain observations as well as optical phenomena.

II. THE VERTICAL DISTRIBUTION OF PRESSURE AND TEMPERATURE IN THE ATMOSPHERE AND THE EFFECT OF VERTICAL DISPLACEMENTS ON SUCH DISTRIBUTIONS

Up to this point certain general relationships between temperature, pressure, and density in the atmosphere have been derived, but no consideration of the variation of these quantities with change of elevation has been undertaken. Since atmospheric pressure is a consequence only of the weight of the overlying atmosphere, it is obvious that change of elevation effects change of atmospheric pressure and consequently of density. Once the relationship between changes of pressure and elevation has been derived, all the relationships which have been obtained involving atmospheric pressure may be expressed in terms of elevation instead.

The hydrostatic equation.—If the atmosphere is to remain in equilibrium in a state of complete rest it is necessary that every external force acting on any portion of it shall be exactly balanced by the prevailing pressure gradient. When this state prevails the atmosphere is said to be in static equilibrium. Since the only external force acting on the atmosphere is that of gravity, it follows that in static equilibrium there is no horizontal pressure gradient, and that the vertical gradient is just balanced by the force of gravity, or $\frac{1}{\rho}\frac{dp}{dz} = -g$, $dp = -\rho g dz$. This is known as the hydrostatic equation. It expresses the fact that the difference in atmospheric pressure between two levels is equal to the weight of the intervening air column. If Figure 76 represents a vertical

column of unit cross section through the atmosphere, and s_2 and s_1 separated by the elevation differential dz are two isobaric surfaces at which the pressure is p_2 and p_1 then the difference between the pressures p_2 and p_1 is the difference in the weight of the air column above s_2 and s_1. This is the weight of the column dz, which is $\rho g dz$. Hence $p_2 - p_1 = \rho g dz$, the hydrostatic equation.

The integration of this equation gives the relation between pressure and elevation. Actually the integration cannot be carried out without making some assumption regarding the variation of ρ or T in the vertical. If it is assumed that ρ remains constant with elevation, a condition

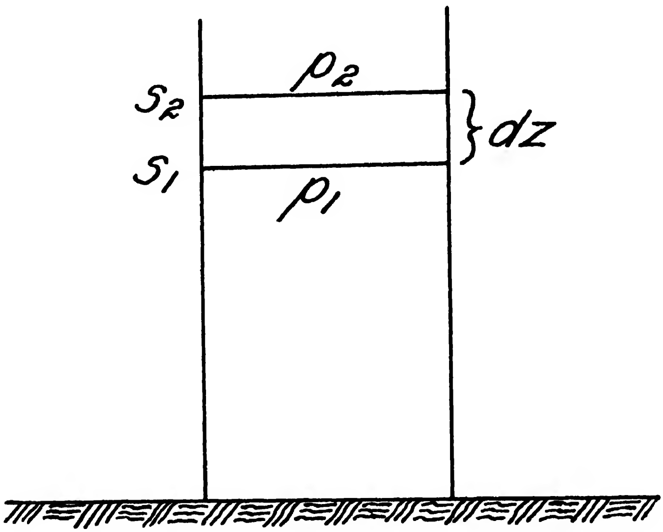

FIG. 76.—Relation between pressure and elevation.

which never obtains generally in the atmosphere, the integration gives $p = p_0 - \rho_0 g z$. If $p = 0$, the condition at the upper limit of the atmosphere, and if the substitution $\rho_0 = \frac{p_0}{RT_0}$ is made, then $z = \frac{RT_0}{g} = H$, the so-called *height of the homogeneous atmosphere*. Evidently, for the same gravity acceleration g, which is assumed constant for all approximate calculations, H depends only on the temperature T_0 prevailing at the base of the atmosphere and is quite independent of the pressure. Since the density is to remain constant throughout the atmosphere, it is evident that the temperature must decrease rapidly with elevation to absolute zero at the upper limit where $p = 0$. For a surface temperature of

$$T_0 = 273°A,\ H = \frac{287 \times 273}{9.805} = 7991 \text{ m.}$$

If the assumption is made that the temperature decrease with elevation is linear and equal to a, the hydrostatic equation becomes

$$dp = -\rho g dz = -\frac{gp}{RT}\,dz, \quad \text{or} \quad \frac{dp}{p} = -\frac{gdz}{R(T_0 - az)},$$

whence

$$\log\frac{p}{p_0} = -\frac{g}{R}\int_0^z \frac{dz}{(T_0 - az)} = \frac{g}{Ra}\log\frac{T_0 - az}{T_0},$$

or

$$\frac{p}{p_0} = \left(\frac{T_0 - az}{T_0}\right)^{\frac{g}{Ra}} = \left(\frac{T}{T_0}\right)^{\frac{g}{Ra}},$$

a relationship resembling Poisson's equation. Substituting

$$\frac{p}{p_0} = \frac{\rho RT}{\rho_0 RT_0} = \frac{\rho T}{\rho_0 T_0}, \text{ it follows that } \frac{\rho}{\rho_0} = \left(\frac{T}{T_0}\right)^{\frac{g}{Ra}-1}$$

In the preceding paragraph the assumption was made that $\frac{\rho}{\rho_0} = 1$. From the equation just derived it is evident that this condition is satisfied by the linear temperature lapse rate given by

$$\frac{g}{Ra} - 1 = 0, \quad \text{or} \quad a = \frac{g}{R} = \frac{9.805}{287} = 0.0342 \text{ degrees per meter},$$

or 3.42 degrees per 100 m. elevation. This lapse rate gives a temperature of 0° A at the elevation $H = 7991$ m., starting from a ground temperature of 273° A. Much more important is the fact that this lapse rate represents the limit of mechanical equilibrium in the atmosphere. For a decrease of temperature with elevation in excess of 3.42 C. per 100 m. the density must increase with elevation $\left(\frac{\rho}{\rho_0} > 1\right)$, and an overturning of the existing stratification must quickly occur.

The usual assumption made in the integration of the hydrostatic equation in the form

$$\int_{p_0}^{p} \frac{dp}{p} = -\frac{g}{R}\int_0^z \frac{dz}{T}$$

is that the mean value of T between $z=0$ and $z=z$, or T_m, may be set equal to the arithmetical mean of the observed temperatures T_0 and T. Then $\log\frac{p}{p_0} = -\frac{gz}{RT_m}$, or $p = p_0 e^{-\frac{gz}{RT_m}}$, where the mean temperature T_m of the air column from 0 to z is set equal to $\frac{T_0 + T}{2}$, which is to a certain extent justified by the normally linear character of the vertical lapse rate.

The lower the observation level z, or the more numerous the intermediate observations of T, the more closely can the actual value of T_m be approximated, and the more accurately will the above formula indicate the actual decrease of pressure with elevation.

The hypsometric equation.—The integration of the hydrostatic equation makes the determination of elevation differences possible directly from the measurement of pressure differences. The difficulty lies in the determination of the distribution of the density ρ in the intervening air column. For the accurate determination of elevation, it is necessary to take into account the variation of ρ not only with T, but also with the vapor content of the air, since water vapor is lighter than dry air. Furthermore, the slight variations in the gravity constant g with latitude and elevation should not be overlooked. Humphreys [5] gives the complete derivation of the formula and expresses the final result in the form

$$h = 7991 \log\left(\frac{p_0}{p}\right) \frac{T_m}{273} \frac{1}{(1-0.605q_m)} \frac{G}{g_m},$$

where G is the value of g at sea-level and latitude 45°, or $9.80665 \dfrac{m}{\sec^2}$, and the subscript m refers to mean values between $z=0$ and $z=h$. The quantity $g_m = G\psi(l)f\left(\dfrac{d}{R}\right)$, where ψ and f are known functions of the latitude l and the ratio $\dfrac{d}{R}$ of elevation above sea-level to the radius of the earth at the point of observation. As usual, q is specific humidity, the mean values T_m and q_m are to be determined again by observation, the arithmetical means of the values at $z=0$ and $z=h$ being used in the absence of intermediate values. In the case that both h and the pressure p are accurately known, the equation can be solved for p_0. This is the problem of the reduction of station barometer readings to sea-level values, for comparative purposes. In the case of elevated stations on land the problem is complicated by the necessity of assuming mean values of T and q for an air column which does not even exist. In this connection it is interesting to note the work of Meisinger.[6] In attempting to reduce station barometer readings over the central and eastern U. S. to upper air levels instead of sea-levels, he found that the correlation between surface wind direction and the mean temperature of the air column up to the one and two kilometer levels at individual stations was so close that a quick reduction, and one sufficiently accurate for synoptic purposes, of surface pressure to pressure at the one and two kilometer levels can be carried out from the observation of surface conditions alone. He made numerous applications of these methods to particular situations, and in

this way was able to explain many features appearing on the surface synoptic charts in terms of air currents aloft whose existence one would not suspect from the surface pressure distribution.

Although the condition of static equilibrium on which the hypsometric formula is based is never completely satisfied in the atmosphere, ordinary atmospheric conditions depart so slightly from the relationship expressed by the hydrostatic equation that it is used quite generally to determine the vertical pressure distribution in the atmosphere.

Temperature changes in a rising unsaturated air mass.—The hydrostatic equation, because it expresses the change of pressure with elevation, offers the means of determining the change of temperature of an air mass resulting from its vertical displacement. Within a dry air mass which is rising and expanding adiabatically, $C_p \frac{dT}{T} = AR \frac{dp}{p}$ (Poisson's Equation). Since $dp = -\rho g dz$, and $p' = p = \rho R T$ (where the primed characters refer to the atmosphere in which the rising air mass is imbedded), it follows that

$$C_p \frac{dT}{T} = -AR \frac{\rho' g dz}{\rho R T}, \quad \text{or} \quad \frac{dT}{dz} = -\frac{Ag}{C_p} \cdot \frac{\rho'}{\rho}.$$

The constant term

$$\frac{Ag}{C_p} = \frac{g}{427 \cdot 0 \times 0.2375} = .00986 \frac{g}{G}.$$

Actually the ratio $\frac{g}{G}$ varies by only 0.3 per cent from unity, and so it is usually omitted. The ratio $\frac{\rho'}{\rho}$ depends on the humidity and temperature differences between the rising air mass and the surrounding air. Even under the assumption that the rising air mass is perfectly dry, the humidity difference seldom if ever causes a density difference of as much as 1 per cent. In the atmosphere the rising air mass is never dry, but as long as it remains unsaturated it will satisfy approximately Poisson's equation for dry air (see page 135). If humidity differences are neglected, then since $p = p'$, it follows that $\frac{\rho'}{\rho} = \frac{T}{T'}$; and consequently the equation $\frac{dT}{dz} = -\frac{Ag}{C_p} \cdot \frac{T}{T'}$ gives approximately the rate of cooling of any unsaturated air mass rising adiabatically through the atmosphere. Usually the ratio $\frac{T}{T'}$ is close to unity, and therefore $\frac{dT}{dz} = -\frac{Ag}{C_p} = -\gamma$, or about 0°.01 C. per m. or 1° per 100 m. A vertical lapse rate of this amount in the atmosphere is known as the *dry adiabatic lapse rate.* In the case of

a prevailing lapse rate of this amount, a rising unsaturated air mass is everywhere in equilibrium with its surroundings.

Stability.—By Archimedes' principle an air mass surrounded by air of different density will be subject to a vertically accelerating force, positive upwards, of $g(\rho'-\rho)$ per unit volume, or an acceleration of $g\left(\frac{\rho'-\rho}{\rho}\right)\frac{\text{m}}{\text{sec}^2}$. Neglecting possible small humidity differences, $\frac{\rho'}{\rho}=\frac{T}{T'}$ and the vertical acceleration may be expressed $g\left(\frac{T-T'}{T'}\right)$. Consequently, relatively warm air tends to rise, relatively cold to sink. If an unsaturated air mass of the same temperature as its surroundings is displaced upward, rising adiabatically, it must cool at the dry adiabatic lapse rate of approximately 1° C. per 100 m. change of elevation. If the prevailing lapse rate in the surrounding air is less than the dry adiabatic, then the rising air mass becomes continuously colder than its surroundings, and consequently denser. Therefore it has a tendency to sink back to its original position. If displaced downward, it becomes warmer and lighter than the surrounding air, and this time tends to rise up to its original position. Therefore, if the prevailing lapse rate in the atmosphere is less than the dry adiabatic, the atmosphere is said to be in ***stable equilibrium***. If the prevailing lapse rate is greater than the dry adiabatic, the displaced air mass becomes warmer than its surroundings if displaced upwards, colder if displaced downwards, and therefore tends automatically to be displaced further in the same direction in which it is started. In this case the atmosphere is said to be in ***unstable equilibrium***. It is to be noted, however, that as long as the density of the atmosphere decreases with elevation (a condition which obtains, as shown on page 142, until the lapse rate becomes 3°.42 per 100 m.), an initial impulse is necessary to start the displacement. When the lapse rate is exactly the dry adiabatic the displaced air mass is continuously at the same temperature as its surroundings so that it tends to remain at that point to which it has been brought. This condition is called ***neutral equilibrium***. Since for adiabatic changes the potential temperature remains constant, the adiabatic lapse rate $\frac{dT}{dz}=-\gamma$ may also be expressed $\frac{d\theta}{dz}=0$. Hence the conditions for stable, neutral, and unstable equilibrium in the atmosphere are $\frac{d\theta}{dz}>0$, $\frac{d\theta}{dz}=0$, $\frac{d\theta}{dz}<0$, and in general $\frac{d\theta}{dz}=\frac{\theta}{T}(\gamma-\alpha)$, where α is the prevailing lapse rate of the actual temperature, or $-\frac{dT}{dz}$. This general relationship

between $\frac{d\theta}{dz}$ and $\frac{dT}{dz}$ may be derived directly from the definition of potential temperature, $\theta = T\left(\frac{p_0}{p}\right)^{\frac{AR}{C_p}}$, by differentiation with respect to z and the substitution $\frac{dp}{dz} = -\rho g = -\frac{pg}{RT}$. Evidently at $\gamma = \alpha$, $\frac{d\theta}{dz} = 0$, and as α passes from values less than γ to those greater than γ, $\frac{d\theta}{dz}$ changes from positive to negative values.

Temperature changes in a rising saturated air mass.—It was found (see page 138) that the heat of condensation liberated in the process of cooling by adiabatic expansion decreases the rate of cooling. Therefore the rate of cooling in a rising saturated air mass is less than that in a dry mass in proportion to the amount of condensation taking place. By the differentiation of the approximate relationship

$$C_p \log T - AR \log p + 0.623\,\frac{re_m}{pT} = \text{constant}$$

with respect to the variables e, p, and T and the elimination of dp with the hydrostatic equation $dp = -\frac{pgdz}{RT}$, the change of temperature with elevation is found to be

$$\frac{dT}{dz} = -\frac{\left(A + \frac{qr}{RT}\right)g}{C_p - \frac{qr}{T} + \frac{qr}{e}\,\frac{de_m}{dT}},$$

where $q = 0.623\,\frac{e_m}{p}$, and where e_m and $\frac{de_m}{dT}$ are known functions of T. Evidently as q approaches 0, or T becomes very low $\frac{dT}{dz}$ approaches the dry adiabatic value $-\frac{AG}{C_p}$, but for the large values of T occurring in the tropics it may become less than 40 per cent of the dry adiabatic. A prevailing lapse rate in the atmosphere such that a rising saturated air mass will remain at the temperature of its surroundings is called the saturation adiabatic lapse rate. Evidently such a lapse rate is not constant, or linear, but with increasing elevation approaches the linear dry adiabatic rate. If the prevailing lapse rate in the atmosphere is between the dry and the saturation adiabatic rates the atmosphere is in stable equilibrium as defined above. However, a rising saturated air mass at the temperature of its surroundings must continuously become warmer than the surrounding air, and therefore continue to rise. Hence the atmosphere is not stable

for saturated air which has been brought to an elevation at which its temperature is in the slightest degree in excess of its surroundings. Such an atmosphere is called *conditionally stable,* the condition for stability being that no vertical displacements shall occur sufficient to result in saturation and the transport of the initially cooler saturated mass to an elevation where it will become warmer than its surroundings. Figure 77 represents

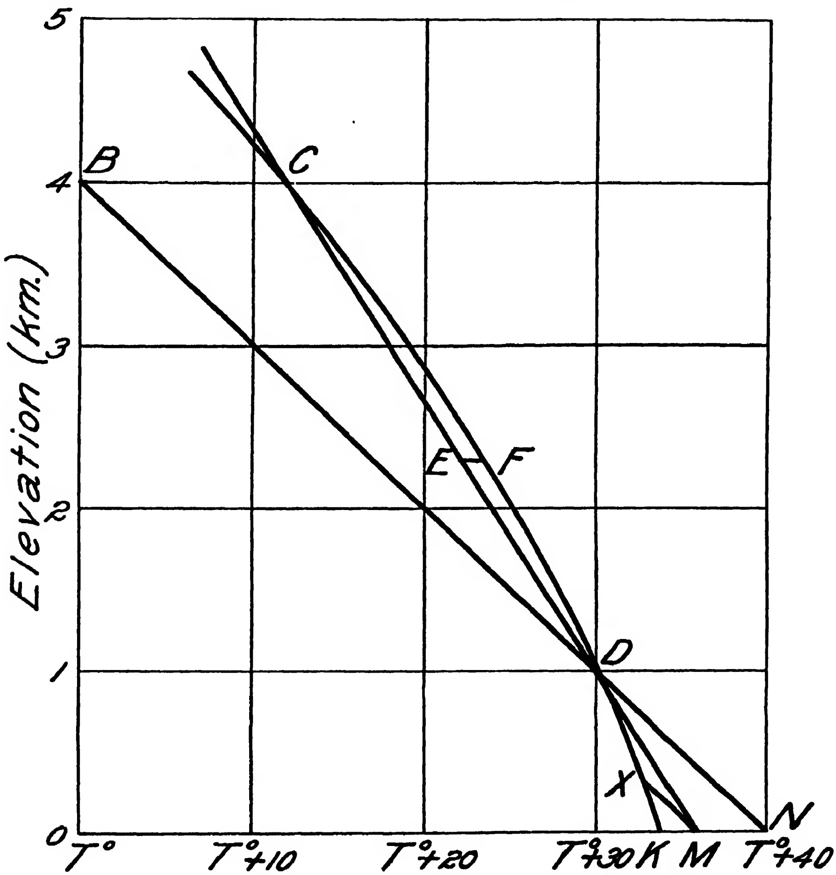

FIG. 77.—Conditional equilibrium.

a conditionally stable equilibrium. *NDB* gives the dry adiabatic lapse rate, *MDEC* the prevailing lapse rate, and *KDFC* the saturation adiabat for air having the temperature and humidity which prevails at *M*. If such a mass is forced to rise, it cools at the dry adiabatic rate, reaching saturation at *X*. From this point it cools as indicated by the saturation adiabat, reaching the same temperature as its surroundings at *D*. At some point *F*, its temperature difference relative to its surroundings reaches a maxi-

mum positive value EF, and above F the difference decreases, becoming negative above C. The atmosphere is in a state of conditional stability from the surface up to the elevation F, for only in this region does the actual lapse rate lie between the dry and the saturation adiabatic rates.

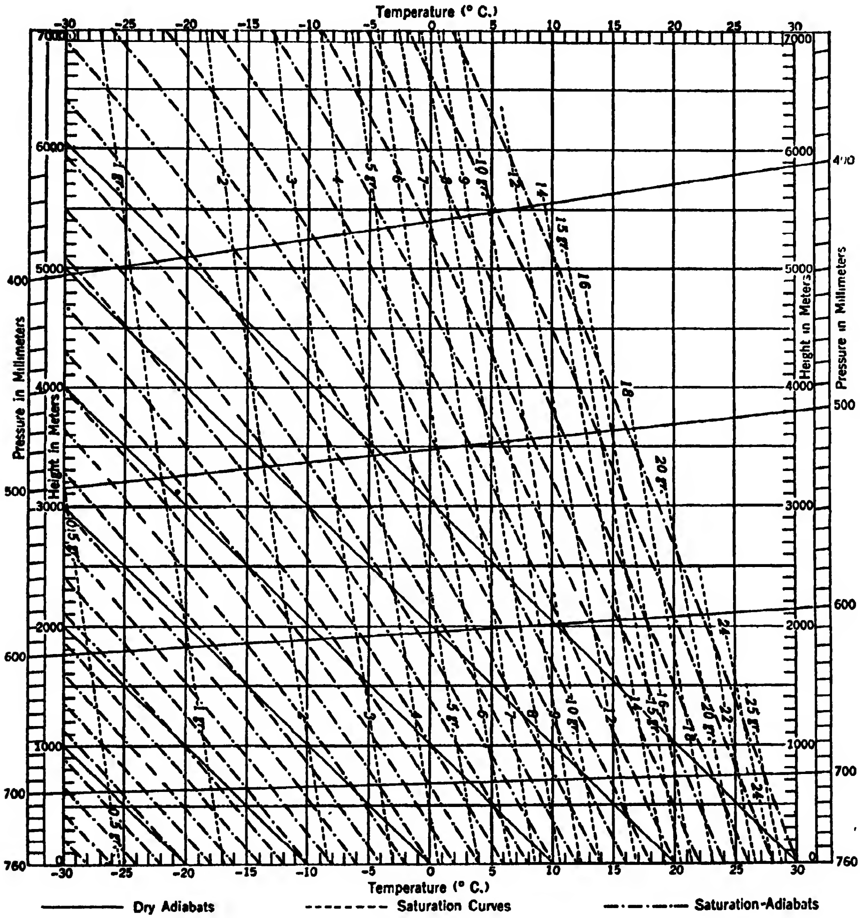

Fig. 78.—Adiabatic diagram (Neuhoff).

The Neuhoff diagram.—Two diagrams are in general use to illustrate the relationships under consideration, namely the Neuhoff diagram and the tephigram. The Neuhoff diagram represents graphically the relations existing between temperature, pressure, elevation and water vapor in adiabatic atmospheric changes. Figure 78, taken from Abbe's Mechanics of the Earth's Atmosphere, opposite page 494 (Smithsonian Misc. Coll., vol. 51, no. 4 (1910)), illustrates the method. A more recent

recalculation of this diagram including corrections of the heat of sublimation at low temperatures has been made by Fjeldstad.[2] His paper contains a diagram on a large scale representing over a wide range of temperature and pressure the results of his computations.

In Figure 78, temperatures are plotted as abscissas, elevations as ordinates. The light unbroken nearly horizontal lines rising slightly from left to right represent the isobaric surfaces, assuming a uniform ground pressure of 760 mm. Their slope is due to the increasing density, and consequent lowering of the isobaric surfaces, resulting from a decrease of the temperature. The light unbroken lines running diagonally upward from right to left represent the decrease of temperature with elevation, or with pressure, of the dry adiabatically expanding air body. They are therefore lines of equal potential temperature. The dot and dash curved lines which approach the slope of the dry adiabats in their upper portions represent the adiabatic expansion of saturated air bodies. The increase of potential temperature indicated in the expanding saturated air mass by the crossing of the lines of equal potential temperature from left to right is due to the supply of heat by condensation. Since these changes are nearly pseudoadiabatic, in case the rising saturated air mass descends again, it follows the dry adiabat instead of the wet, remaining potentially warmer than it was during the ascent. The fine broken lines give the number of grams of water vapor per kg. of dry air necessary to effect saturation at the given T and p. If the temperature and relative humidity at the ground (760 mm.) are given for a body of air, from the saturation curves the approximate number of grams of water per kg. of dry air may be directly determined. Hence if an air mass expands and cools following a particular dry adiabat, the temperature at which condensation begins may be determined directly from the diagram by noting the point at which the dry adiabat intersects the proper saturation curve for the particular quantity of moisture found to belong to the rising air mass. This obviates the necessity of using Magnus' formula or tables to determine the saturation temperature. In the case of descending motion or compression the temperature changes in the sinking air mass follow the course of the dry adiabats.

The tephigram.—The Neuhoff diagram permits of the presentation of the thermodynamic state of the atmosphere. By plotting temperature against pressure on the Neuhoff diagram for increasing elevations, and noting the specific humidities, it is possible to determine at once the prevailing state of convective equilibrium, and in cases of instability or conditional stability to determine the displacement of each air layer necessary to effect neutral equilibrium. But the Neuhoff diagram does not

permit of a determination of the actual amount of energy available in the existing atmospheric stratification for the production of convective turbulence. In order to obtain an energy diagram of atmospheric conditions and to make possible a graphical computation of the energy available for convective turbulence, Sir Napier Shaw developed the tephigram.

As coordinates of the tephigram the abscissæ represent absolute temperature T, the ordinates represent entropy of unit mass of dry air. In plotting points on the diagram to represent conditions at different points in the atmosphere, the presence of water vapor is not considered. Only when condensation has occurred does the moisture make its in-

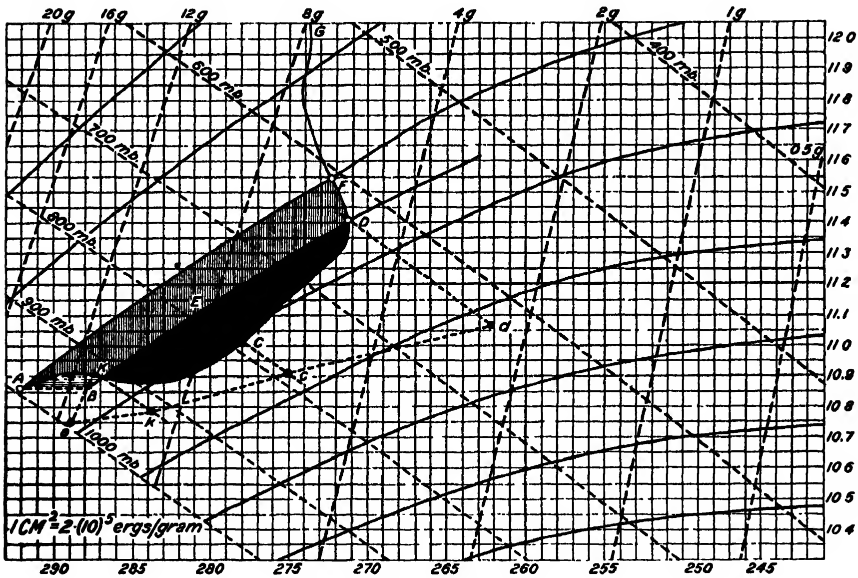

FIG. 79.—The tephigram.

fluence detectable by the increased entropy of unit mass of dry air caused by the liberation of heat of condensation. From the relationship $S = C_p \log \theta + \text{constant}$ (see page 136), it follows that the horizontal isentropic lines of the tephigram are also isotherms of potential temperature, and represent the dry adiabatic lapse rate in the atmosphere. Vertical lines represent isothermalcy.

From the relationship $S = C_p \log \frac{T}{T_0} - AR \log \frac{p}{p_0}$ (see page 135), it follows that for any arbitrarily chosen point of reference T_0 and p_0, the entropy S of unit air mass is quantitatively determined by its T and p. Shaw uses the values $T_0 = 100°A$ and $p_0 = 1000$ mb. in fixing his entropy

values. Isobars are plotted in the temperature-entropy system of co-ordinates simply by holding p constant in the above equation. They are indicated in Figure 79 by the broken lines which rise from right to left. In case of saturation, the entropy of a rising air mass is no longer constant as is the case for dry adiabatic changes, but increases by the amount of the heat of condensation set free. Hence the saturation adiabatic lines in the temperature entropy system are represented by the full lines (see Figure 79) which rise from left to right, approaching at low pressures the horizontal position of the dry adiabats. They are computed from the relationship for saturated air (see page 138)

$$C_p \log T - AR \log p + 0.623 \frac{re_m}{pT} = K,$$

which corresponds to the Poisson relationship for dry air where the third term takes into account, approximately, the heat of condensation. e_m is the saturation vapor pressure at the temperature T, r the heat of condensation. It is also convenient to have plotted for reference on the temperature-entropy diagram the lines of specific humidity for saturated air. Lines of equal specific humidity q at saturation are plotted from the relationship (see page 137) $q = \frac{e_m}{1.605p - 0.605e_m}$, where e_m is a function of T alone. They are represented in Figure 79 by the broken lines inclined slightly to the right of the vertical.

The actual plotting of the tephigram is a simple matter with the observational material from an upper air sounding at hand, though there are corrections to be applied to the observed temperature, pressure, and humidity readings before they are ready for plotting. From this material points are plotted on the temperature entropy diagram for chosen elevations, and the smooth curve passed through these points is the tephigram. In Figure 79 $AKCDFG$ is such a curve. The degree of saturation may be represented by plotting the corresponding dew point curve (depegram). For each observation point on the tephigram a corresponding point on the same isobar but at the dew point temperature as determined from the observed humidity may be plotted. $akcd$ represents a depegram determined in this way for the points $AKCD$ of the tephigram. The length of the lines Aa, Kk, etc., represent graphically the extent to which the atmosphere falls short of saturation at the corresponding points.

The real value of the tephigram is better utilized by another method of taking into account the presence of moisture. If unit mass of air at the point A were saturated, and rose adiabatically, its entropy and potential temperature would change as indicated by the saturation adiabat AF. At every point between A and F the entropy of unit mass of dry air in

the rising saturated mass would be greater than that of the surrounding atmosphere at the same pressure, therefore its temperature must be higher, and it must be in a state of convective instability. The integral $\int T ds$ measures the increase of energy of any air mass under changing conditions. For unit air mass this is represented on the temperature-entropy diagram by the total area between the curve followed by the changing air mass and the line of zero potential temperature T. The difference between the area under AF and that under the tephigram $AKCDF$, *i. e.*, the area $AFDCKA$, represents the total energy of the rising unit mass of dry air under saturation conditions which is available for convective turbulence. Such an area lying above the tephigram is said to be a positive area, for it represents positive energy available for the production of convective instability. If the tephigram had followed a course above the saturation adiabat the area would have been negative. The condition would have been one of convective stability as far as unit air mass at the point represented by A is concerned, *i. e.*, energy would have to be supplied to the air mass to effect convection. A similar representation of the energy available for convective turbulence for unit air mass at any other point on the tephigram may be made in the same way. For practical use only the point or points on the tephigram which will give the maximum positive area are chosen for representation. In general, if the tephigram falls below the slope of the saturation adiabats the condition is unstable for saturated air; if it falls below the horizontal dry adiabats, it is unstable for unsaturated air as well; that is, there will be positive areas between even the dry adiabats and the tephigram. Usually the air at any point such as A is not saturated, but must cool adiabatically to its saturation temperature at some point B before it follows the saturation adiabat. In that case the positive area may entirely disappear, or at least be greatly reduced, as in the figure $KECD$, and there may appear a negative area ABK representing an initial stable layer which must be penetrated before the rising air mass will be in a state of convective instability. Such a situation represents a condition of conditional stability in the atmosphere. (See page 147.)

A mathematical demonstration, due to C. G. Rossby, of the equality between available energy and the area between the tephigram and the dry or saturation adiabats may be found in a paper by C. M. Alvord and R. H. Smith.[47] Figure 79 is taken from the same paper. On the scale on which it is drawn, 1 sq. cm. area represents an energy of 2.10^5 ergs. per gram of dry air. These authors have made a detailed study of tephigrams in a large number of summer situations, particularly with a view to the forecasting of local convectional thunderstorms. They came to the

conclusion that in general, thunderstorm conditions become evident in the appearance of large positive areas on the tephigram, at least six hours before the occurrence of thunderstorms, and that for that reason a tephigram based on a morning upper air sounding is the most reliable indication there is as to the probability of thunderstorms later in the day. They think that frontal disturbances, *i. e.*, disturbances connected with atmospheric discontinuities (see page 188 *et seq.*), also are detectable on the tephigram at least six hours before the approach of the front to the observing station.

Observed temperature gradients in the atmosphere.—The distribution of temperature at the earth's surface is fairly well known, but the increasing numbers of upper air observations by kite, aeroplane, and sounding balloons (see Chapter IV) still leave a very incomplete picture of upper air conditions. However, certain facts stand out clearly. Most striking of these is the existence of the so-called *stratosphere,* or isothermal region, in the upper portion of the atmosphere. In this region the temperature is nearly constant, or perhaps increases slightly with elevation, as far as reliable observations can be obtained. The elevation and temperature of the base of the stratosphere vary with latitude, season, and the presence of anticyclonic high or cyclonic low pressure at the earth's surface. On the average it rises from an elevation of some 6 km. and a temperature of perhaps $-45°$ C. at the poles, to a probable 17 km. and temperature of $-80°$ over the equator.* Hence in the substratosphere regions the normal poleward temperature gradient existing at the earth's surface is reversed, a fact which is of importance for the general atmospheric circulation. In general the stratosphere is somewhat higher and warmer in summer than in winter, but the difference in temperature between summer and winter is less than in the lower atmosphere. Also, the stratosphere is higher, sometimes by as much as two or even three kms., and colder over anticyclones than over cyclones. An explanation of the reason for the existence of the stratosphere and the equatorward temperature gradient can be offered only after a consideration of radiation phenomena and the general atmospheric circulation.

In the portion of the atmosphere below the stratosphere, called the troposphere, the mean vertical temperature gradient is found to be about 6° C. per km. In general, it is slightly greater in equatorial regions, slightly less in polar regions, and slightly greater in the upper troposphere than near the ground. Thus for the most part it is slightly in excess of the saturation adiabatic lapse rate, but considerably less than the dry adiabatic. Through the first two kms. it is noticeably steeper in summer

* See Fig. 75, Chap. IV, p. 125.

than in winter, the result of insolational heating and radiational cooling. The significance of these lapse rates lies in the prevailing stability or increase of potential temperature with elevation which they indicate in the atmosphere. It is not convective equilibrium but the equilibrium of saturation convection which prevails, hence even in the troposphere the atmosphere is in a state of thermal stratification. Therefore a moving air mass must continue in horizontal motion in its own stratum unless forced to another level, where it can remain only by a change of its potential temperature. It follows that vertical movements in the atmosphere are distinctly the exception, rather than the rule, and to be postulated in the explanation of atmospheric phenomena only when their existence may be plausibly accounted for or actually observed. Local convection currents may follow the increase of potential temperature of a low air stratum by insolational heating, as in the case of the local thundershower, but the widespread ascending currents which occur in connection with the cyclone are initiated by convergent air flow near the ground, and maintained at least in part by the heat of condensation aloft. Obviously no convective action can penetrate very far into the stratosphere. At the low temperatures prevailing in the upper troposphere and in the stratosphere the amount of condensation taking place is very slight. It is vertical motion in the lower troposphere which gives most of the measured precipitation.

III. EFFECT OF VERTICAL MOVEMENTS IN THE ATMOSPHERE ON THE TEMPERATURE LAPSE RATE

In the last section the effect of vertical displacements on the temperature and pressure within small isolated air masses was considered. The present discussion is concerned with the effect on the temperature distribution throughout large portions of the atmosphere 1) of the vertical transport of entire atmospheric strata without overturning or mixing, and 2) of widespread turbulent mixing.

Large scale or adiabatic movements.—In general the vertical movements occurring in the atmosphere are in the nature of either 1) rather extensive, uniformly moving currents of the convective type in which the rising or sinking air masses are so large that the changes taking place may be treated as adiabatic, or 2) small irregular and disordered motions, usually referred to as *turbulence,* in which the individual elements of the movement very rapidly assume the properties of their surroundings, by conduction and mixing. These two types of vertical air movement work quite different effects on the prevailing lapse rate in the atmosphere. In the case of the large scale currents, the changes taking place are those

occurring within the ascending or descending current. It is assumed that the surrounding atmosphere remains unchanged, and further that the general ascending or descending motion does not extend throughout the atmosphere, but that individual pressure changes occur throughout the vertically moving column in proportion to the change of elevation, as given by the hydrostatic equation. Given the thin layer of an air column of cross section q at the isobaric surface p (see Figure 80) at which there prevails the lapse rate $\frac{dT}{dz}$, it is required to find the lapse rate $\frac{dT'}{dz'}$ after the air stratum has suffered an adiabatic displacement downward to the

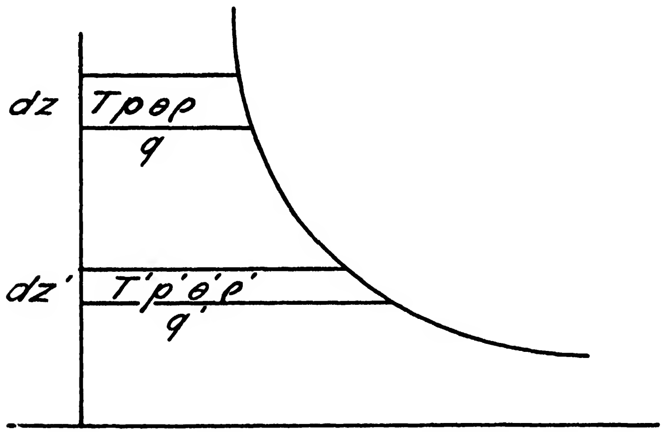

FIG. 80.—Vertical displacement with horizontal spreading of an air column.

isobaric surface p' and expanded to the cross section q'. This problem is best solved by working with potential temperature θ, for the changes being adiabatic θ remains constant.

At the initial position the prevailing potential temperature lapse rate is given by

$$\frac{d\theta}{dz} = \frac{\theta}{T}\left(\gamma + \frac{dT}{dz}\right),$$

where γ is the dry adiabatic lapse rate. (See page 145.) In the final position the corresponding lapse rate is

$$\frac{d\theta'}{dz'} = \frac{\theta'}{T'}\left(\gamma + \frac{dT'}{dz'}\right).$$

Since the changes are adiabatic, $\theta = 6'$ and $d\theta = d\theta'$, hence

$$\frac{d\theta'}{dz'} = \frac{d\theta}{dz'} = \frac{d\theta}{dz} \cdot \frac{dz}{dz'} = \frac{\theta}{T'}\left(\gamma + \frac{dT'}{dz'}\right).$$

Furthermore,

$$\frac{dz}{dz'} = \frac{\rho' q'}{\rho q} = \frac{p' T q'}{p T' q}, \text{ for } \rho = \frac{p}{RT}.$$

Therefore

$$\frac{d\theta'}{dz'} = \frac{d\theta}{dz} \cdot \frac{p'Tq'}{pT'q} = \frac{\theta}{T}\left(\gamma + \frac{dT}{dz}\right)\frac{p'Tq'}{pT'q} = \frac{\theta}{T'}\left(\gamma + \frac{dT'}{dz'}\right)$$

or

$$\left(\gamma + \frac{dT}{dz}\right)\frac{p'Tq'}{pT'q} = \frac{T}{T'}\left(\gamma + \frac{dT'}{dz'}\right)$$

or

$$\frac{dT'}{dz'} = \frac{p'q'}{pq}\left(\gamma + \frac{dT}{dz}\right) - \gamma, \text{ the required lapse rate.}$$

If

$$\frac{dT}{dz} < \gamma, \text{ then when } \frac{q'p'}{qp} > 1, \ \frac{dT'}{dz'} < \frac{dT}{dz},$$

and when

$$\frac{q'p'}{qp} < 1, \frac{dT'}{dz'} > \frac{dT}{dz}.$$

In general the condition $\frac{q'p'}{qp} < 1$ holds in ascending motion, and $\frac{q'p'}{qp} > 1$ in descending motion. Therefore if initially stable there will be a tendency to a decreasing lapse rate in a sinking air mass, and increasing one in a rising mass. The ratio $\frac{q'}{q}$ becomes markedly greater than unity as a rising air column approaches the stratosphere, and as a sinking column approaches the ground, because here the vertical movement cannot continue, so that a general spreading out must take place. Therefore in the upper troposphere ascending motion may lead to a decreased lapse rate or even inversion, in spite of the fact that $\frac{p'}{p}$ remains a proper fraction. In descending motion near the ground both $\frac{p'}{p}$ and $\frac{q'}{q}$ are greater than unity, so that a marked decrease or inversion of the lapse rate may be expected. This is generally given as the explanation of the extreme thermal stability and usual development of one or more inversions in anticyclones consisting of a gradually sinking and spreading cold air mass of northerly origin.

Small scale or turbulent movements.—Wherever there is a shearing motion in the normally horizontal atmospheric movements, numerous small eddies or turbulent motions are generated. This takes place primarily at the ground, when the wind blows over the rough surface. The rougher the surface and the stronger the wind, the greater the turbulence produced. But it also takes place at higher elevations espe-

cially at surfaces of marked discontinuity of wind velocity. Richardson [8] first attacked theoretically the problem of the formation and dissipation of turbulent energy in the atmosphere. Evidently the source of this turbulent energy is to be found in the kinetic energy of adjacent air currents of different velocities. Consequently the more rapid the increase of velocity with elevation, or the greater the shearing motion, the greater will be the rate of production of turbulent energy. On the other hand, the turbulent energy produced in this manner is dissipated in two ways 1) by transformation into heat energy through molecular friction, and 2) by increasing the internal or potential energy of the turbulent air mass through vertical displacements against a stable stratification, *i. e.,* by carrying potentially colder air up from below, potentially warmer down from above. Richardson showed that usually the second influence is so much greater than the first that the first may be entirely neglected, so that whether a given layer in the atmosphere is undergoing a net gain or loss of turbulent energy depends upon the lapse rate and the vertical distribution of wind velocity. He was able to formulate a definite quantitative relationship, or criterion, by which to determine whether production or dissipation of turbulent energy is taking place. By this criterion to every value of the lapse rate there corresponds a critical value of the increase of wind velocity with elevation. If the actual increase is in excess of this critical value, turbulence or eddy energy is being produced, if it is less than the critical value, eddy energy is being dissipated. For the lapse rates normally prevailing on clear days Richardson finds the critical value of wind velocity increase with elevation to be about 1 m/sec. per 100 m.

If the lapse rate exceeds the dry adiabatic, the mechanical turbulence passes over into the large scale convective movement, for now the unstable stratification becomes a source of turbulent energy. Therefore properly any discussion of mechanical turbulence applies only to a stable atmosphere. Actually the only difference between the effects of convective and those of mechanical turbulence is due to the difference in the elementary vertical motions. In the former case they are so large that the vertically moving currents do not mix to form a homogeneous atmosphere whose properties may be discussed; on the contrary the properties of the ascending and descending currents must be considered separately from those of the surrounding atmosphere. For the small scale currents of mechanical turbulence it is assumed that they rapidly mingle with their surroundings, imparting to the atmosphere at the point of mixing the properties brought from their place of origin. Assuming that there is no general vertical motion, mechanical turbulence consists of numerous small eddies or minute ascending and descending currents, such that across any

horizontal surface the total movement of mass upwards must equal the movement downwards. The exchange of mass across unit horizontal surface in unit time measures the degree of turbulence. For this quantity the German name *austausch* is now generally used. Since in this process small air masses from one level are constantly moving to higher and lower levels, mingling there, it follows that the tendency of turbulence is to effect uniform vertical distribution in the atmosphere of any conservative property which varies with elevation. Conservative properties which may be effectively treated in this way include heat, water vapor, momentum, and solid impurities in the air. W. Schmidt[7] has made many interesting quantitative applications of this principle in air and water. The simplest quantitative expression for austausch and the turbulent transfer of atmospheric properties may be obtained as follows: Let s be the measure of any atmospheric property per unit mass of atmosphere, so that to a mass m there belongs the quantity ms, and let s vary with elevation. Then every particle moving downward across any horizontal surface in the atmosphere at which the concentration of the property is s transports downwards the quantity of s given by $m\left(s+\frac{\partial s}{\partial z}dz\right)$, where dz represents the vertical distance of the source of m above the surface, and the total downward flow is given by $\Sigma m\left(s+\frac{\partial s}{\partial z}dz\right)$. Similarly, the total upward flow of s is given by $\Sigma m\left(s-\frac{\partial s}{\partial z}dz\right)$. Since there is to be no general vertical motion of the atmosphere, the Σm moving downward must equal the Σm moving upward. Therefore the net flow of s across the surface, being the difference between the upward and the downward flow, becomes $-\Sigma m dz\,\frac{\partial s}{\partial z}$, where Σm is now the total mass movement across the surface. Across unit surface and in unit time the upward flow of s is

$$-\frac{\Sigma m dz}{ft}\frac{\partial s}{\partial z} \quad \text{or} \quad -A\,\frac{\partial s}{\partial z},$$

where f is the total area of the surface considered, t the total time of the observed net flow, and A is the austausch coefficient, or simply austausch. This last, it will be noticed, depends not only on the exchange of mass, but also on the distance from which this mass has moved. Thus the flow of the property s depends on the vertical gradient $\frac{\partial s}{\partial z}$, and on the austausch or turbulent activity. The austausch itself depends upon the presence of shearing motions in the atmosphere and the vertical lapse rate. As long as there is any horizontal velocity whatsoever there is bound to

be more or less turbulence. The presence of a temperature inversion is an almost complete shield against the upward spread of turbulence.

The change of the property s per unit volume due to turbulence, or $\frac{\partial \rho s}{\partial t}$, is equal to the difference between the net flow of s across the upper surface and across the lower surface of the unit volume. If the horizontal plane where the concentration is s passes through the center of the unit volume, then

$$\frac{\partial \rho s}{\partial t} = -\left[\left(A\frac{\partial s}{\partial z} - \frac{\partial}{\partial z}A\frac{\partial s}{\partial z}\frac{1}{2}\right) - \left(A\frac{\partial s}{\partial z} + \frac{\partial}{\partial z}A\frac{\partial s}{\partial z}\frac{1}{2}\right)\right]$$

or

$$\frac{\partial \rho s}{\partial t} = + \frac{\partial}{\partial z}A\frac{\partial s}{\partial z}.$$

If it be assumed with W. Schmidt[7] that A is nearly constant with elevation, then $\frac{\partial \rho s}{\partial t} = +A\frac{\partial^2 s}{\partial z^2}$. Actually, as Richardson[8] has pointed out, A is not nearly constant, but increases many fold in the first few hundred meters above the ground, due to the checking of vertical motions by the ground surface. Since ρ may be considered constant through the comparatively small portion of the atmosphere to which turbulence discussions are usually applied, the rate of change of s per unit mass is

$$\frac{\partial s}{\partial t} = + \frac{A}{\rho}\frac{\partial^2 s}{\partial z^2}.$$

In considering the effect of turbulence on the vertical lapse rate, it must be remembered that the heat content of the atmosphere is measured by the potential temperature rather than the actual temperature. Hence the direction of the vertical flow of heat due to turbulence depends on the sign of $\frac{d\theta}{dz}$, and since this quantity is positive for a stable lapse rate, mechanical turbulence normally effects a flow of heat downward to the ground. The actual vertical heat flow is given by $s = -\sigma A\frac{\partial \theta}{\partial z}$, where σ is the specific heat of the turbulent air. Only in the case of a superadiabatic lapse rate $\left(\frac{d\theta}{dz} < 0\right)$ does convective turbulence effect an upward flow of heat. Therefore the tendency of all turbulence, as long as condensation does not occur, is to bring about a uniform potential temperature distribution, or an adiabatic lapse rate, in the atmosphere. Since the gain or loss of heat in any unit volume of the atmosphere depends on $\frac{\partial^2 \theta}{\partial z^2}$, in any portion of the atmosphere in which the lapse rate decreases with

elevation $\left(\frac{\partial^2\theta}{\partial z^2}\text{ positive}\right)$ there is a net gain of heat taking place due to turbulence, where it increases, a net loss of heat occurs. The flow of heat normally taking place downwards towards the ground W. Schmidt[7] has shown to be quite considerable. Since the normal lapse rate is nearly the saturation adiabatic, evidently as the result of saturation convection equilibrium, the source of this downward turbulent flow of heat must be a part of the heat of condensation carried up as uncondensed water vapor. As wind increases turbulence, other things being equal the tendency of an increasing wind is to steepen the vertical lapse rate. From instances where $\frac{\partial s}{\partial z}$ and the actual flow of the property s, or $\frac{\partial^2 s}{\partial z^2}$ and the change of s per unit volume, can be observed and measured, numerical values of the austausch coefficient A may be determined observationally. W. Schmidt has made such observations under varied conditions, and has found the mean value for the air layer just above the ground to be about 50 $cm^{-1}g$ sec^{-1}, being about 59 in summer and 44 in winter.

Water vapor normally decreases with elevation, hence the normal effect of turbulence is to carry moisture upwards, whereas it carries heat downwards. Therefore the upper portion of a stable air layer tends to become colder and moister under the action of turbulence, a process which sometimes leads to supersaturation and the formation of low stratus.

The vertical transport of horizontal momentum by turbulence is very important owing to its effect on the vertical distribution of wind velocities. This question will be considered in its proper place, in the discussion of frictional drag between air layers. (See page 184, *et seq.*)

Heat may be transported by conduction, or the spread of molecular activity, in a manner very similar to that of turbulent activity. This transport is expressed by a very similar equation, actual instead of potential temperature determining the flow, but the coefficient of molecular austausch is less than $\frac{1}{10,000}$ that of active vertical turbulence. Hence heat conduction in the air is an extremely slow process, and is usually neglected in the study of meteorological problems. Besides mechanical and convective turbulence, including the transport of water vapor, the important heat transporting processes for meteorology are radiation and horizontal austausch, or advection.

IV. RADIATION

Solar radiation.—Observations show that the sun is for all practical purposes the sole source of the heat which maintains the temperature of the earth's surface and the atmosphere. The heat received from the moon

and stars is entirely negligible, while the temperature lapse rate and coefficient of conduction observed in the earth's outer crust indicate a flow of heat from the interior sufficient to increase the temperature of the earth's surface only by about 0°.1 C. The effect of this heat on the temperature distribution in the atmosphere is quite insignificant. Since the sun is the sole source of the heat present in the atmosphere, and since the unequal supply of this heat to different portions of the atmosphere is the underlying cause of the atmospheric circulation, it is of the first importance to ascertain how and where this heat is supplied. Measurements of the solar radiation reduced to extra-atmospheric intensity and mean solar distance show that the sun radiates heat energy at an almost constant rate, which at the distance of the earth amounts to slightly less than two cal. (1.94; see Chapter III, page 51) per min. and per sq. cm. normal to the incident radiation. This quantity is called the *solar constant* I_0. If there were no atmosphere, the intensity of solar radiation received per unit area of level ground surface, neglecting the slight variations of the solar distance due to the eccentricity of the earth's orbit (see Chapter III, p. 35), would be $I_0 \sin \alpha$, where α is the angular elevation of the sun above the horizon. α varies with season, latitude, and time of day. With the sun on the equator the radiation incident per level square cm. per day in the absence of the atmosphere would be $458.4 I_0 \cos \psi$, where ψ is the latitude. Thus a maximum of heat is received at the equator, falling off to zero at the poles. Actually, the inclination of the earth's axis to the plane of the ecliptic effects an apparent seasonal movement of the sun in the heavens, so that during half the year either pole receives solar radiation which reaches a maximum at the summer solstice of about 25 per cent more per day than the maximum daily amount received at the equator when the sun is overhead. The total amounts of incident radiation in kg. cal. per cm.^{2} level surface in the course of a year at different latitudes are given by Defant [9] as follows:

TOTAL INCIDENT SOLAR ENERGY PER SQUARE CENTIMETER

Latitude	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°
Summer half-year	160.5	169	173.5	173	169	160.5	149	139	134	133
Winter half-year	160.5	147	129.5	109	84	58.5	33	13	4	0
Annual	321	316	303	282	253	219	182	152	138	133

It will be noticed from these figures that in the course of a year about 40 per cent as much solar energy is received at the poles as at the equator,

although a smaller percentage of this incident radiation will be effective in heating the earth's surface at the poles, due to the greater reflection from snow and ice surfaces. It is also evident that the incident solar energy falls off fastest from the fiftieth to the sixtieth latitude circles. This corresponds to the fact that it is between these latitudes that the poleward temperature gradient at the earth's surface is greatest, and consequently the air movement, or storminess, and the advective poleward heat transport are a maximum (see page 168).

Of the total incident solar energy distributed as indicated in the above table, only that part actually absorbed by the atmosphere or ground is effective in heating the earth's surface. All energy so absorbed can be re-emitted to space only as the long-wave terrestrial heat radiation, corresponding to the low temperature (relative to the sun) of the earth's radiating surface. That portion of the incident solar radiation not actually absorbed and transformed to the long-wave terrestrial radiation is returned to space unchanged, without having produced the slightest heating effect. The fraction of the total incident solar energy so returned is generally estimated at about 37 per cent, and is a physical characteristic of the earth called its albedo. This return to space of unchanged solar radiation is effected in part by reflection in the atmosphere, especially from clouds, in part by a scattering to space by the gas molecules of the atmosphere, and in part by reflection from the earth's surface, especially by ice, snow, and water. Of the remaining 63 per cent which is absorbed and makes its presence felt as heat, a small part, usually estimated at about 10 per cent, is absorbed in the atmosphere, and the remainder makes its way either directly or as diffuse scattered sky light to the earth's surface * where it is absorbed. Since the earth's temperature remains approximately constant, eventually all the effective incident solar energy must be radiated back to space as long-wave terrestrial radiation. The earth's atmosphere, due primarily to its water vapor content and perhaps to a lesser extent to its ozone and carbon dioxide, is highly absorptive of this terrestrial radiation. Whereas it is assumed to absorb only 10 per cent or less of the incoming short-wave solar radiation, it is believed to take up 90 per cent or more of the outgoing terrestrial radiation. Since the radiation absorbed in an air stratum is reradiated equally upwards and downwards, 45 per cent of the outgoing terrestrial radiation is returned by the atmosphere, to be reradiated and half returned again, etc. Actually the amount of this absorption depends upon the amount of moisture in the atmosphere above the radiating point. According to figures given by Defant,[10] the radiation returned to the earth's surface

* See Chap. III, Figs. 12 to 17, p. 48 and 49.

by the atmosphere ranges normally from $\frac{2}{3}$ to $\frac{3}{4}$ of what the total radiation from that surface to space would be at the same prevailing temperature if there were no interference by the atmosphere. For high mountain stations this ratio is reduced to from $\frac{1}{3}$ to $\frac{1}{2}$. Clearly the effect of the atmosphere, increasingly marked the greater the water content, is to let through the incoming solar radiation and to intercept the outgoing terrestrial radiation so that the surface temperature of the earth is considerably higher than it would otherwise be. This is the so-called *greenhouse effect* of the atmosphere.

Radiation equilibrium.—Evidently each thin layer of the atmosphere is traversed by two radiation currents, an upward current A and a downward current B. The upward current consists of a long wave length portion, A_2 the terrestrial radiation, and a much smaller short wave portion, A_1, which represents the unchanged solar radiation reflected or scattered back to space from lower levels. Similarly, the downward current B consists of a larger short wave portion B_1, the direct solar radiation and the diffuse scattered solar radiation, or skylight, and a smaller long wave portion B_2, the return radiation to the earth from the upper atmosphere. If the coefficient of absorption of the thin atmospheric layer dm for the short and long wave radiation may be taken as k_1 and k_2, respectively, then the thin layer absorbs the different amounts k_1B_1dm, k_1A_1dm, k_2B_2dm and k_2A_2dm. If E_1dm and E_2dm are the amounts of the short and long wave radiation which would be emitted in each direction by the thin atmospheric layer at its own temperature if a perfect radiator, then it actually does emit from each side the amounts k_1E_1dm, and k_2E_2dm. This follows from Kirchhoff's law, that the ratio of the emission to the absorption of any body depends upon the temperature only and is numerically equal to the emission of a perfectly black body at the same temperature and wave length. The absorptive thickness of the atmospheric layer is dm, m representing the absorptive element, probably chiefly water vapor, in the atmosphere. The changes in the intensity of the currents A and B in passing through dm are given by the differential equations

$$\frac{dB}{dm} = -k_1B_1 - k_2B_2 + k_1E_1 + k_2E_2$$

$$\frac{dA}{dm} = k_1A_1 + k_2A_2 - k_1E_1 - k_2E_2,$$

dm being positive downwards. The condition of radiation equilibrium is that absorption and emission of radiation shall everywhere be equal, and hence the temperature distribution constant. This condition applied to the atmospheric layer dm gives

$$2(k_1E_1 + k_2E_2) = k_1B_1 + k_2B_2 + k_1A_1 + k_2A_2.$$

This is Emden's [11] condition for radiation equilibrium in the atmosphere assuming selective absorption. A_1 and E_1 are insignificant and may be neglected, then $2k_2E_2=k_1B_1+k_2B_2+k_2A_2$ where $\frac{E_2dm}{k_2}=\sigma T^4$, on the assumption that there is a constant coefficient of absorption or radiation k_2 for the absorptive element of the atmosphere in the rather restricted portion of the spectrum in which the energy of its emitted radiation is concentrated at the prevailing temperature T. σ is the proportionality constant of Stefan's law of black-body radiation.

In order to determine the temperature distribution which would prevail in the atmosphere under radiation equilibrium, Emden assumed that the effective solar radiation (63 per cent of the total incident) is reradiated equally from all parts of the earth's surface. He also assumed that the entire absorption of radiation by the atmosphere is due to water vapor, so that B_1 is diminished in passing through the thickness m of this element in accordance with the formula for the extinction of radiation, $B_1=Ie^{-\int_0^m k_1 dm}$, where I is the value of B_1 before entering the atmosphere. Emden assumes further that the distribution of water vapor in the atmosphere is that given by Hann's empirical formula $f=f_0 10^{-\frac{h}{6000}}$, where h is the elevation. Finally, the assumption that 90 per cent of the solar radiation and only 10 per cent of the terrestrial radiation passes through the atmosphere without being absorbed gives a value of the ratio $\frac{k_1}{k_2}=\frac{1}{23}$. Under these assumptions Emden finds that the temperature of the atmosphere is given at any point by

$$T^4=\tau^4\frac{k_1+k_2}{2k_1k_2}[k_2-(k_2-k_1)e^{-k_1m}],$$

where $\tau=254°$ A., the temperature the earth radiating as a black body would require to get rid of the heat actually received from the sun. Since already at 11 km. m (moisture) becomes vanishingly small, above this elevation

$$T=254°\sqrt[4]{\frac{1+\frac{k_1}{k_2}}{2}},\quad\text{or for }\frac{k_1}{k_2}=\frac{1}{23},\quad T=-57°\text{ C.},$$

the temperature of the stratosphere, which agrees very well with the observed mean value. The limiting value of T for $\frac{k_1}{k_2}=0$ is $-59°$ C., the stratosphere radiation temperature which would prevail if none of the incoming radiation B_1 were absorbed in the atmosphere. Emden

assumes that at very high elevations, where there is practically no water vapor, k_1 and k_2 lose their characteristic different values, the ratio approaching unity, and T therefore increasing again to $-19°$ C. At the earth's surface Emden's radiation equilibrium formula for T gives a temperature of $+15°.8$ C. This temperature falls off very rapidly through the first few kms. and thereafter gradually approaches the isothermal value above 11 kms. That there exists no such steep lapse rate through the first three kms. Emden accounts for by the action of convection and condensation, which continue until a lapse rate barely in excess of the saturation adiabatic is established from the ground upwards approximately to the stratosphere, where true radiation equilibrium begins. This explanation of the troposphere as being in convective, and the stratosphere in radiation equilibrium, seems to account very well for the observed average conditions, but since it is based on an assumption of equal disbursement of heat from all portions of the earth, it naturally does not account for the surprising latitudinal variations in the properties of the stratosphere. Furthermore, as Hergesell[12] first pointed out, the radiation equilibrium temperature distribution which Emden finds for the troposphere is quite incorrect. Hergesell made the point that the distribution of water vapor in the atmosphere assumed by Emden would lead to enormous supersaturation, a relative humidity of 6,000 per cent at 6 kilometers, if the radiation equilibrium temperatures calculated by him actually prevailed in the atmosphere. Immediately most of the vapor would condense and fall out, the absorption of radiation would be greatly reduced and therefore the prevailing temperature decreased, then more condensation would take place, and this process would continue until isothermalcy at the stratosphere temperature,

$$T = 254° \sqrt[4]{\frac{1+\frac{k_1}{k_2}}{2}},$$

prevailed from the ground up. Hence Emden's radiation equilibrium alone would permit of no troposphere whatsoever.

Hergesell went further, and, on the assumption that only water vapor in the atmosphere absorbs and radiates, he made use of the vertical humidity and temperature distribution actually obtained from sounding balloon ascents over Lindenberg in Europe and over Batavia in the tropics to calculate the radiation absorbed and emitted at every elevation in the atmosphere. He found that above about 12 km. the amount of water vapor, and therefore of absorption and emission, becomes negligibly small. But he found that in the troposphere the emission greatly

exceeds the absorption, the difference reaching a maximum at from 3 to 4 kms. Since conditions are in the long run stationary, he assumed that the heat balance is maintained in intermediate levels by advection and vertical convection with condensation. Probably from 3 to 4 kms. is actually the level at which the greatest amount of condensation of water vapor occurs. This is just another way of stating again that it is convective and not radiation equilibrium which exists in the troposphere.

A. Ångström has carried out in southern California and Algeria a series of direct measurements of the long wave downward radiation (Gegenstrahlung) at different elevations. He finds,[13] like Hergesell, that the amount of moisture in the atmosphere becomes negligible before the stratosphere is reached, but he does not find that the Gegenstrahlung vanishes with the water vapor, as Hergesell's assumption that water vapor is the only significant absorbing element in the atmosphere would imply. On the contrary, Ångström observes a variation of the Gegenstrahlung up to 5 km. which if extrapolated to greater elevations with a negligible amount of water vapor would approach a constant value of about 40 per cent of the value at the ground. This amount should be observed from 8 kms up. This can only mean that some element present especially in the upper atmosphere, possibly ozone, absorbs and emits a considerable amount of radiation. For the radiation equilibrium of the stratosphere this means that in the expression

$$T = 254^\circ \sqrt[4]{\frac{1 + \frac{k_1}{k_2}}{2}},$$

the ratio $\frac{k_1}{k_2}$ instead of approaching unity above the region of appreciable water vapor, as Emden assumed, probably approaches the particular value characteristic of the absorbing element throughout the stratosphere. Therefore T, instead of approaching 254° A. in the upper stratosphere, probably remains nearly constant at an appreciably lower value. This result of Ångström's observations also modifies Hergesell's picture of the balance of heat in the troposphere, as Hergesell himself points out in closing his discussion. This increase over Hergesell's estimate of the downward stream of long wave length radiation having its source in the upper stratosphere increases the amount of radiation absorbed by the water vapor in the lower atmosphere. This in turn decreases Hergesell's calculated disparity between absorption and emission of radiation throughout the troposphere, and consequently necessitates the postulation of a smaller degree of convective and advective activity in the troposphere in order to maintain the existing balance of heat.

Horizontal advection and the stratosphere.—The entire discussion of the temperature distribution in the atmosphere has to this point been based on the assumption of equal distribution over the earth's surface of the heat received by the terrestrial absorption of solar radiation. Consequently the results obtained have indicated the mean vertical temperature distribution in the atmosphere for all latitudes. R Mügge [14] has applied Emden's formula for the stratosphere radiation equilibrium temperature in the form $\sigma T_\psi^4 = \frac{1}{2} I_\psi \left(1 + \frac{k_1}{k_2}\right)$ to determine the stratosphere temperature at the latitude ψ under the assumption that the solar radiation, $I_\psi = I_E \cos \psi$ is absorbed at latitude ψ. I_E is the mean effective solar radiation received per minute per sq. cm. of surface and overlying atmosphere at the equator. σ is the proportionality factor of Stefan's law. In other words, Mügge assumes the distribution of solar radiation which would obtain if the sun remained always at the equator and all the effective radiation passed directly to the earth without deviation in the atmosphere. He assumes an albedo of 43 per cent. Under these assumptions he finds the stratosphere temperature over the equator to be $-50°$ C., falling off to absolute zero over the poles where no solar radiation is received. This poleward temperature gradient is qualitatively though not quantitatively what would be expected. It remains to explain the actually observed reverse gradient. Mügge points out that one might assume a much greater albedo than 43 per cent over the equator where there occurs so much convectional condensation and cloudiness, but the cold stratosphere cannot be explained in this way without necessitating an equally cold troposphere, and this does not exist. There remains the possibility of the advectional transport of heat from equatorial to polar regions in an amount sufficient to maintain the observed stratosphere temperatures. Mügge adds a term to the stratosphere temperature formula, writing $\sigma T_\psi^4 = \frac{1}{2} I_\psi \left(1 + \frac{k_1}{k_2}\right) - \frac{1}{2} R_\psi$ where R_ψ represents the mean advectional transport of heat in calories per minute from or to each square centimeter column of the troposphere at latitude ψ. Where R is positive the insolation exceeds the radiation to space, the difference being carried off by advection; where R is negative, the radiation to space exceeds the insolation, the difference R being supplied by advection. Mügge calculated the value R_ψ and the latitudinal flow of heat S_ψ, which must be present in the troposphere to give the actually observed stratosphere temperatures T_ψ. He finds that at latitude 35°, R_ψ goes from positive to negative values, or the transition from regions of excess insolation to those of excess radiation is made. Due to the decreasing circumference of the latitude circles, the flow across each cm. of a circle of latitude, here called S_ψ, reaches a

maximum at latitude 45°; while on the other hand, due to the decreasing height of the troposphere the flow across each sq. cm. of the plane normal to the flow reaches its maximum at latitude 55°. At this latitude the flow must amount to 50 $\frac{\text{g. cal.}}{\text{min. cm.}^2}$ throughout the troposphere, which is a very large value. However, this is the region of the maximum poleward temperature gradient and of maximum storminess, which indicates that it must also be the region of the maximum intensity of advectional heat transport. Furthermore, the computed value of 50 $\frac{\text{g. cal.}}{\text{min. cm.}^2}$ agrees very well with the value of the horizontal austausch (latitudinal interchange of air masses) as computed by Defant [15] for this latitude. The net result of this heat transport is that more heat is radiated to space per sq. cm. in the polar regions than in the tropical. At the pole itself the total radiated energy is the transported amount $R_{90°}$, which is found to be even in excess of the total absorbed incident solar radiation per cm.^{2} at the equator. That the actually observed stratosphere temperatures indicate such a heat flow, and that the actually existing interzonal circulation is sufficient to effect it, there seems to be little doubt. The question as to why a latitudinal transport of heat should exist in the troposphere of such vigor as to reverse completely the natural poleward temperature gradient in the stratosphere can be answered only after an elementary consideration of thermodynamic equilibrium. The decrease in elevation of the stratosphere from equator to pole is evidently a consequence of the great decrease in the temperature difference between the ground and the stratosphere, and therefore of the vertical distance through which the convectional equilibrium and mean saturation lapse rate of the troposphere can persist.

Actually, the seasonal journey of the sun north and south of the equator somewhat modifies the above representation of the atmospheric temperature distribution. Whichever hemisphere is favored by the sun's presence receives a considerably greater supply of insolational energy than indicated above, especially in the polar regions. Therefore in general the horizontal austausch and latitudinal transport of heat are greatly diminished, the troposphere becomes warmer, and the stratosphere becomes warmer in the same proportion, hence somewhat less warmer in actual number of degrees. Therefore, although the stratosphere becomes warmer in summer, the difference in temperature between it and the earth's surface increases, and on the assumption of a constant lapse rate, its elevation must be somewhat greater in summer, an observed fact which has already been pointed out. In winter the reverse of all these conditions holds. There is a marked deficit in solar insolation, especially in the polar regions, a big increase in horizontal austausch and heat transport, a colder tropo-

sphere and a proportionately colder stratosphere which is at a lower elevation because the actual temperature difference from the ground is lessened.

In connection with the problem of horizontal heat transport and the difference between insolation and radiation, the recent calculations of G. C. Simpson [16] are of great interest. Simpson followed the very difficult method of calculating for the different months the actual amount of heat radiated to space at different latitudes on a basis of surface temperature, stratosphere temperature, cloud surface temperature, and amount of cloudiness. In the same way he calculated the effective solar insolation by months for different latitudes, on the basis of an albedo determined from mean cloudiness, and assigned the large constant value of 0.65 in the ice- and snow-covered polar regions. In spite of the large albedo assumed for polar regions, Simpson finds a much smaller deficit than Mügge of heat in these regions to be supplied by horizontal transport even in the winter. In fact, if his calculations are correct, Defant's determination of the horizontal austausch gives much too large values. The fundamental difference between Simpson's results and Mügge's lies in the much greater value Mügge obtains for the terrestrial radiation to space from the polar regions. It is difficult from Simpson's determination of the distribution of terrestrial radiation to account either for the warm polar stratosphere or for the observed interzonal circulation. Mügge's calculations, on the other hand, are based on the existence of the former, and his results indicate the necessity of the latter. However, to account for a greater radiation to space from the colder earth of the polar regions, it is necessary to assume that the actual surface of effective radiation is higher and colder in the tropics than in the polar regions. This in turn depends on the much greater water vapor content of the tropical atmosphere and the carrying of this element to high elevations by extensive convection.

The explanation of the cold high stratosphere over the large semi-permanent anticyclones, and the reverse over the cyclones, may be in part dynamic, but is doubtless in part similar to that of the equatorial-polar differences. Mügge [14a] has developed this hypothesis at length. The incident solar radiation in the anticyclone is absorbed and warms the troposphere, but is largely carried off in the general anticyclonic outflow in the form of water vapor, and is radiated back to space from the upper condensation levels of the cyclone. The high albedo in the cyclonic region prevents the solar heating of the troposphere there. It seems probable, however, that the dynamic factors are quite as important as the radiational in effecting local variations in elevation and temperature of the stratosphere above cyclones and anticyclones.

V. THE DYNAMICS OF AIR CURRENTS

In the last section it was pointed out that extensive horizontal air movements are necessary for the continued existence of the observed radiation equilibrium in the atmosphere. At this point, then, it becomes necessary to examine the forces which come into play when horizontal air flow exists, and to inquire how the unequal distribution of solar energy gives rise to the compensating currents.

Horizontal pressure gradient.—The condition under which the atmosphere shall everywhere be and remain at rest is that the isobaric surfaces shall everywhere be horizontal. This requires that the isothermal and isosteric (equal specific volume) surfaces shall also be horizontal. It has been pointed out that the departures from this condition occurring in the atmosphere are so slight that the hydrostatic equation holds approximately at all times. The normally prevailing horizontal pressure gradients are from 1/10,000 to 1/100,000 of the vertical. Although these gradients are neglected in assuming the application of the hydrostatic equation in considerations of vertical equilibrium, as soon as the prime consideration is one of horizontal air movements, or ordinary winds, the horizontal pressure gradient is of the first importance. The horizontal pressure gradient depends upon the slope of the isobaric surfaces, or the closeness of the successive intersections of the horizontal plane by the isobaric surfaces (isobars). The horizontal gradient in the direction x, or $\frac{\partial p}{\partial x}$, is usually expressed as mm. of Hg. per degree of latitude.

Just as it was found that the vertical pressure gradient $\frac{\partial p}{\partial z}$ causes an upward force $-\frac{1}{\rho}\frac{\partial p}{\partial z}$ per unit mass of air, which in the state of vertical equilibrium is just balanced by the downward force of gravity g, so the horizontal gradient $\frac{\partial p}{\partial x}$ causes a force of $-\frac{1}{\rho}\frac{\partial p}{\partial x}$ in the direction of x positive per unit mass. Since there is generally no external force acting in the atmosphere to balance a horizontal pressure gradient, the appearance of such a gradient is always balanced by an acceleration of the air in the direction of the gradient. If u and v are velocities in the direction of x and y positive (east and north) then the simple equations of rectilinear horizontal motion are

$$\frac{du}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} \text{ and } \frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y}.$$

Actually as soon as there exist velocities u and v other accelerations

come into play which will be considered presently. The corresponding equation for vertical acceleration is $\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$, w usually being negligible. In static equilibrium every term in these three equations is 0 except $\frac{1}{\rho}\frac{\partial p}{\partial z}$ and g.

The circulation principle.—Let AB (Figure 81) represent a portion

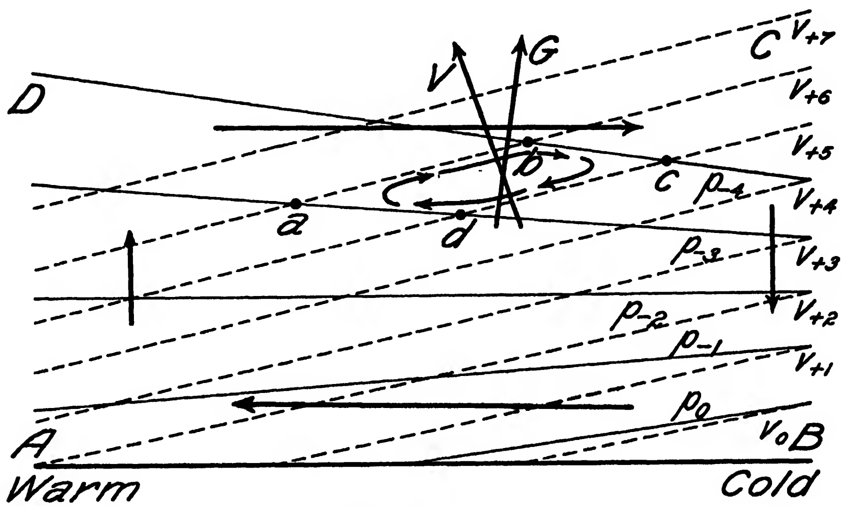

FIG. 81.—Solenoids and circulation.

of the earth's surface above which the air is at rest, or the isobars and isosteres run parallel to the ground. If unequal heating occurs, for instance if heat is added at A, and removed at some point C above B, the atmosphere above A is heated and expands, while that above B is cooled and contracts. It follows that the isosteric surfaces will now slope from B to A, while the expansion of the atmosphere above A lifts the isobaric surfaces so that at upper levels they slope increasingly towards C. The resulting horizontal pressure gradient leads to a flow of air from D to C, which, in turn, causes at the ground a reduced pressure at A, an increased pressure at B, and therefore a gradient and flow of air from B to A. The warmed air at A rises and the cooled air at C sinks, maintaining the outflow from D and B and therefore the closed circulation $DCBAD$. It is evident that the acceleration of this circulation depends upon the difference in slope of the isobaric and isosteric surfaces. If V represents the direction at any point normal to the isosteric surfaces of increasing specific volume, and G the direction at the same point normal to the

isobaric surfaces of decreasing pressure, then the tendency at that point is to an increasing circulation such that V will coincide more nearly with G, as indicated by the curved arrow in Figure 81. The air column included between two consecutive isobaric and isosteric surfaces is known as a unit solenoid, or simply a solenoid. The parallelogram *abcd* is the cross-section of such a solenoid. Since the acceleration of the circulation *ADCBA* depends upon the relative slope of the isobaric and isosteric surfaces, it might be expected that the number of solenoids inclosed by this curve would give a measure of the acceleration of the circulation about the curve. In order to find the quantitative relationship existing between solenoids and circulation it is necessary to consider briefly just what is meant by circulation and its acceleration.

If s represents any connected chain of fluid particles forming a closed circuit, the circulation about the closed curve s in the fluid is defined as $C=\int^{s} U_t ds$, where ds is an element of the curve s and U_t is the component of the total velocity U tangent to the curve s. This may also be written

$$C=\int^{s} U_t ds=\int^{s}\left(u\frac{dx}{ds}+v\frac{dy}{ds}+w\frac{dz}{ds}\right)ds=\int^{s}(udx+vdy+wdz).$$

The acceleration of the circulation

$$=\frac{dC}{dt}=\int^{s}\left(\frac{du}{dt}dx+\frac{dv}{dt}dy+\frac{dw}{dt}dz\right),$$

for

$$\frac{d}{dt}\int^{s}(udx+vdy+wdz)=\int^{s}\frac{d}{dt}(udx+vdy+wdz)$$

$$=\int^{s}\left(\frac{du}{dt}dx+\frac{dv}{dt}dy+\frac{dw}{dt}dz\right)+\left(u\frac{d}{dt}(dx)+v\frac{d}{dt}(dy)+w\frac{d}{dt}(dz)\right),$$

but

$$\int^{s}\left(u\frac{d}{dt}(dx)+v\frac{d}{dt}(dy)+w\frac{d}{dt}(dz)\right)=\int^{s}(udu+vdv+wdw)=0$$

for $(udu+vdv+wdw)$ is a perfect differential, and s is taken to be a closed curve. If only the pressure gradients and the external force of gravity are considered,

$$\frac{du}{dt}=-\frac{1}{\rho}\frac{\partial p}{\partial x},\quad \frac{dv}{dt}=-\frac{1}{\rho}\frac{\partial p}{\partial y},$$

and

$$\frac{dw}{dt}=-g-\frac{1}{\rho}\frac{\partial p}{\partial z}.$$

Hence

$$\frac{dc}{dt} = \int^{s} - \frac{1}{\rho}\left(\frac{\partial p}{\partial x}\,dx + \frac{\partial p}{\partial y}\,dy + \frac{\partial p}{\partial z}\,dz\right) - g\,dz$$

or

$$\frac{dc}{dt} = -\int^{s} \frac{1}{\rho}\,dp - \int^{s} g\,dz = -\int^{s} v\,dp$$

because, the external force of gravity having a potential, $\int g\,dz$ around a closed curve must be 0. The expression $-\int^{s} v\,dp$ represents the work which is done by the pressure forces on unit air mass in passing around s, and this quantity measures the acceleration of the circulation about s.

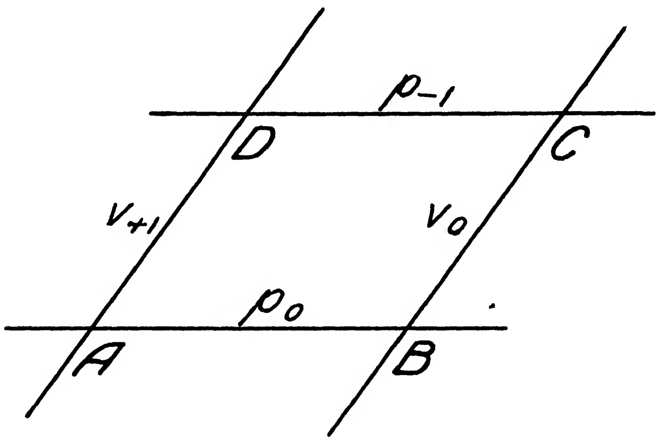

FIG. 82.—Unit solenoid.

If $ADCB$ (Fig. 82) represents a cross section of the unit solenoid inclosed by the isosteric surfaces v_0 and v_{+1} and the isobaric surfaces p_0 and p_{-1}, then the integral $-\int^{ADCBA} v\,dp$ is readily evaluated. Along DC and BA it is 0, for p is constant. Then

$$-\int^{ADCBA} v\,dp = -\left(\int_{p_0}^{p_{-1}} v_{+1}\,dp + \int_{p_{-1}}^{p_0} v_0\,dp\right)$$

$$= -\left[v_{+1}(p_{-1} - p_0) + v_0(p_0 - p_{-1})\right] = -(-v_{+1} + v_0) = +1.$$

Therefore the acceleration of the circulation about a unit solenoid is unity, and in general for any closed curve s,

$$\frac{dc}{dt} = -\int^{s} v\,dp = A,$$

where A is the number of unit solenoids enclosed by s. This is the form of the circulation principle as developed by Bjerknes [17] and first applied

in meteorology by him. For incompressible fluids v is constant, and therefore the circulation is constant (Helmholtz's principle of the constancy of vorticity).

This expression permits of a quantitative determination of the rate of increase of circulation which any given distribution of temperature and pressure will cause. Whenever there occurs a rapid horizontal change of temperature, solenoids are crowded, and circulation or compensating air movements arise. Examples of such circulations are land and sea breezes, thunderstorms, monsoon winds, and even the general circulations of the hemispheres between tropical and polar regions. In fact this principle underlies all natural air movements, *i. e.*, all direct circulations in the atmosphere. There is also the possibility of a forced circulation, or an indirect circulation maintained by the energy of one or two direct circulations. Actually, in the atmosphere, the course of any air current is more or less modified by the earth's rotation, and the energy of motion of any circulation is gradually dissipated by friction. A detailed consideration of these modifying factors will be taken up presently, but at this point it should be pointed out that the acceleration of air movements effected by the presence of solenoids is eventually balanced by the dissipative effect of friction. Thus a steady supply and removal of heat may maintain a steady atmospheric circulation on the principle of a thermodynamic engine. The air is the working body, and the work performed is that of overcoming friction. The energy transformations taking place are fundamentally as follows: Heat energy, supplied either directly by the absorption of solar radiation or indirectly through the condensation of water vapor initially evaporated by the heat of solar radiation, is converted into potential energy of mass distribution, as indicated by the presence of solenoids. This in turn leads to a circulation or transformation of potential energy of mass distribution into kinetic energy of motion, always in a sense to decrease the potential energy or the number of solenoids. The kinetic energy of motion is in turn dissipated by friction, being converted into energy of heat and eventually lost to space as terrestrial radiation.

Of course, the maintenance of a stationary circulation necessitates a cold source as well as a warm source, otherwise the temperature of the entire circulation system must continuously increase. Without friction such a stationary system would be impossible, for without the expenditure of part of the heat received in doing work there must be equality between the heat Q received at the warm source and the heat Q' given up at the cold source and the efficiency of the system, $\frac{Q-Q'}{Q}$, becomes 0. Sandström [18] first pointed out that the energy made available for over-

coming friction or increasing the circulation by the completion of a cycle is equal to $g(h_1-h_2)$, per unit mass of the working body, where (h_1-h_2) represents the difference in the vertical distance at the warm source and the cold source between the two isobaric surfaces which are considered as bounding the closed circulation above and below. This means practically that the strength of the circulation depends upon the elevation of the cold source above the warm source. This is very important in the atmospheric circulation, which is evidently especially favored by the loss of heat to space by radiation from the upper troposphere and elevated land masses in the polar regions (Greenland and the Antarctic continent).

As Bjerknes points out, the fact that the earth is rotating necessitates a certain modification of the circulation theorem if it is to apply to actual atmospheric circulations. Because the earth itself is in rotation and because the atmosphere moves with it, the total circulation around any closed circuit s is really $C=C_r+C_e$, where C_r is the observed circulation relative to the earth and C_e is the circulation relative to fixed axes of reference of the plane of the surface of the rotating earth at the point in question. Any circuit s lying in the plane of rotation of the earth is rotating with the angular velocity ω of the earth itself. It is readily shown that the integration of the tangential velocity due to this rotation around s is $2\omega s'$, where s' is the total area included by the closed circuit. If s does not lie in the plane of rotation of the earth, then the circulation around s due to the earth's rotation is given by $2\omega s''$, where s'' is the projection of the area s' on the plane of the equator. Thus $C=\int^s U_t ds+2\omega s''$, and the acceleration of the circulation relative to the earth becomes

$$\frac{dC_r}{dt}=\frac{dC}{dt}-\frac{dC_e}{dt}=A-2\omega\frac{ds''}{dt},$$

depending not only on the number of solenoids A, which determines the acceleration of the total circulation C, but also on the rate of change of the projection of the area inclosed by s on the plane of the equator. Usually, however, the last term is of little significance and the presence of solenoids remains the important factor in all atmospheric circulations relative to the earth.

The effect of the earth's rotation on horizontal air movements.—Any air current moving freely over the earth's surface is affected in the following several different ways by the rotation of the earth about its axis:

a) *Effect of east-west or west-east velocities on weight.*—Because the earth is rotating from west to east and the atmosphere with it, there is a certain centrifugal force acting on the atmosphere away from the earth's

axis. This force may be resolved into two components, one horizontal, greatest at the poles which tends to make the atmosphere slide up towards the equator, and one vertical, greatest at the equator, which tends to balance to a slight extent the force of gravity. Evidently any movement of the atmosphere with a west-east component increases the angular velocity ω with which that portion of the atmosphere is rotating about the earth's axis, and an east-west component decreases the angular velocity ω. Hence a west wind is acted on by a slightly greater centrifugal force than an east wind, so that the east wind, other things being equal, will have a slight tendency to underrun the west wind and to move poleward. However, this effect is always so slight compared with other factors, especially differences of temperature or density, that it can always be neglected.

b) *Conservation of angular momentum.*—Any body moving freely in an orbit under the influence of only centrally acting forces tends to maintain a constant angular momentum, or $\omega r^2 = C$ (Kepler's law of equal areas). If air masses moving from one parallel of latitude to another followed this law in their rotation about the earth's axis, it is readily seen that extensive latitudinal displacements would of necessity give rise to very great wind velocities, west-east for a northward moving mass, and east-west for a southward moving mass. Since extensive latitudinal displacements of air masses are frequently observed, but wind velocities corresponding to constancy of angular momentum are never observed it must be concluded that other forces are at work which prevent the realization of such velocities. These forces are the pressure and frictional forces, in part skin friction at the earth's surface, but still more the internal friction or mutual drag of air masses of different angular momentum and velocity giving rise to pressure gradients which always oppose the excessive velocities of individual air masses. However, the more or less continuous interzonal exchange of air masses (horizontal austausch) should make itself felt as a constant deficit of angular momentum (relative to that of the rigid earth surface) in southerly latitudes, and a corresponding excess in northerly. This is most marked in the tropics and sub-tropics where the interzonal circulation is most regular (trades and antitrades), possibly accounting in part for the great zone of strong east winds over the equator, winds which fall off and become westerlies further north. (See page 200.)

c) *The horizontal deflective force due to the earth's rotation.*—At either pole of the earth the plane of the earth's surface is perpendicular to the axis of the earth's rotation and therefore rotating with the angular velocity ω of the earth itself. If this rotation be represented in the usual way as a vector of length ω in the direction of the axis of rota-

tion, then evidently the component of this rotation in the plane of the earth's surface at latitude ψ is $\omega \sin \psi$. Hence the rotation of the earth's surface in the plane of the horizon falls off from ω at the poles to zero at the equator. The following simple demonstration of horizontal deflection in the plane of rotation is that given by Humphreys in Physics of the Air.

Suppose an air particle is moving freely over a plane of reference rotating with constant angular velocity ω about P (Figure 83). Assume that at the pole P the particle has the constant linear velocity in space of v which, if the plane of reference were at rest, would bring it in the

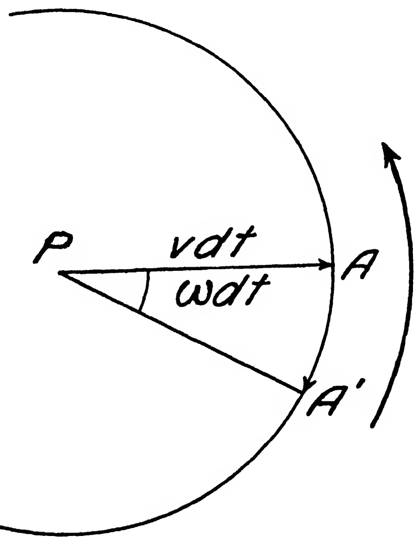

FIG. 83.—Horizontal deflection due to the earth's rotation.

short time dt to the point A. However, due to the rotation of the plane of reference under the moving particle in the direction $A'A$, the particle does not reach A, but rather A', having suffered an apparent displacement relative to the plane of AA'. Actually the particle has been moving freely in the initial direction PA, but the plane has rotated till A' lies in the initial position of A. The apparent acceleration s at right angles to v, which the particle has suffered in the time dt, corresponds to that effected by a force $F = ma$, where from $s = \frac{1}{2}adt^2$ it is evident that $a = \frac{2s}{dt^2}$. Since $s = \omega dt \cdot vdt = \omega vdt^2 = \frac{1}{2}adt^2$, it follows that $a = 2\omega v$, and $F = 2m\omega v$. That is, for unit mass $F = 2\omega v$, the force required per unit mass of freely moving atmosphere to effect the apparent acceleration due to the rotation of the plane of reference. If the plane of reference is the earth's surface,

at latitude ψ where the component of rotation in the plane of the horizon is $\omega \sin \psi$, the apparent horizontal deflecting force due to the earth's rotation is given by $F = 2\omega \sin \psi v$ per unit mass. In general any particle moving freely over the earth's surface suffers an apparent acceleration to the right (left in the southern hemisphere) at right angles to the direction of its motion, of the amount $2\omega v \sin \psi$. The rate of angular deviation of the moving particle is given by

$$\tan^{-1} \frac{s}{vdt} = \tan^{-1} \frac{\omega v \sin \psi dt}{vdt} = \tan^{-1} \omega \sin \psi = \omega \sin \psi,$$

approximately. Thus the rate at which the moving particle changes its direction of motion relative to the rotating earth depends only on its latitude, and not on its velocity. Hence for small scale movements (such that ψ can be assumed constant) a particle moving freely over the rotating earth describes a circle in its movements (circle of inertia) whose diameter is directly proportional to the velocity of the particle, and whose period depends only on the latitude.

The acceleration $a = 2v\omega \sin \psi$ corresponds to the exertion of a force $2mv\omega \sin \psi$, or $2v\omega \sin \psi$ per unit mass of the moving body, which is usually referred to as the deflective force due to the earth's rotation. However, it must be remembered that the force is purely fictitious, and that the acceleration is due to the rotation of the plane of reference under the moving body, and not to the acceleration of the moving body over a fixed plane of reference. But because in ordinary thought all movements on the earth's surface are referred to that surface as if it were at rest, it is convenient to assume that it is so and to supply a fictitious force to account for the relative acceleration of a body moving freely over it. It must also be remembered that this deflective force can never effect an acceleration in the speed of a particle, but only in its direction, for it always acts at right angles to the direction of movement of the particle.

The deflective force due to the earth's rotation must also be taken into account in the equations of motion in the xy (horizontal) plane (see page 170). Since for v positive (north) this force acts in the direction of x positive (east) and for u positive (east) it acts in the direction of y negative (south), the equations of rectilinear motion for unit air mass become

$$\frac{du}{dt} = 2\omega \sin \psi v - \frac{1}{\rho} \frac{\partial p}{\partial x}, \text{ and } \frac{dv}{dt} = -2\omega \sin \psi u - \frac{1}{\rho} \frac{\partial p}{\partial y}.$$

For brevity the quantity $2\omega \sin \psi$ is usually written l, and the equations for rectilinear horizontal motion are written

$$\frac{du}{dt} = lv - \frac{1}{\rho} \frac{\partial p}{\partial x}, \text{ and } \frac{dv}{dt} = -lu - \frac{1}{\rho} \frac{\partial p}{\partial y}.$$

Steady motion.—If the accelerations $\frac{du}{dt}$ and $\frac{dv}{dt}$ are both zero there must be a steady air flow of constant velocity. Such motion is called steady motion. From squaring $lv = \frac{1}{\rho}\frac{\partial p}{\partial x}$ and $lu = -\frac{1}{\rho}\frac{\partial p}{\partial y}$ and adding the results, it is evident that

$$l(u^2+v^2)^{\frac{1}{2}} = \frac{1}{\rho}\left[\left(\frac{\partial p}{\partial x}\right)^2 + \left(\frac{\partial p}{\partial y}\right)^2\right]^{\frac{1}{2}}.$$

But $\sqrt{u^2+v^2} = c$, the total velocity, and

$$\sqrt{\left(\frac{\partial p}{\partial x}\right)^2 + \left(\frac{\partial p}{\partial y}\right)^2} = \frac{\partial p}{\partial n},$$

the horizontal pressure gradient in the direction n normal to the isobars.

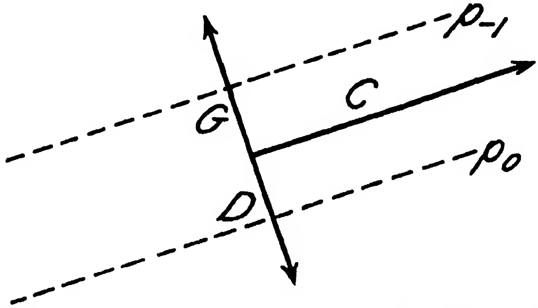

FIG. 84.—Deflective and gradient force equilibrium for gradient rectilinear air flow (Northern Hemisphere).

Hence $lc = \frac{1}{\rho}\frac{\partial p}{\partial n}$, or for rectilinear steady motion the horizontal pressure gradient is just balanced by the deflective force due to the earth's rotation. This is shown schematically in Figure 84. Since the deflecting force D is always vertical to the direction of motion c, it follows that for steady motion the total velocity c must be parallel to the isobars. If it were not, the pressure gradient G, always normal to the isobars, could not exactly balance D, and an acceleration of the velocity c would follow. In the southern hemisphere, where D always acts to the left instead of the right for the same direction of c, the direction of G must also be reversed to maintain equilibrium. Therefore the lower pressure must be found to the right of the direction of c instead of to the left. For steady curvilinear motion the condition that there shall be no acceleration of velocity remains the same as for straight line motion, but in the balance of forces the centrifugal force due to the curvature of the path of the moving air mass must be taken into account. Like the deflecting force,

this force acts always at right angles to the direction of motion c. The force is $\frac{c^2}{r}$ per unit mass of moving air, where r is the radius of curvature of the path of the air movement. Thus the pressure gradient force is balanced by two forces, or $\frac{1}{\rho}\frac{dp}{dn}=lc\pm\frac{c^2}{r}$. It is interesting to note the relative importance of the terms lc and $\frac{c^2}{r}$ under ordinary conditions. For extratropical cyclones of northerly latitudes (l a maximum) for which r is large and c only moderate, the first term outweighs the second four or five times. For the tropical cyclones of southerly latitudes (l a minimum) for which c is very great and r very small, the deflective force becomes quite negligible in comparison to the centrifugal, even falling off to $\frac{1}{40}$ in extreme cases. For anticyclones the deflective force always greatly outweighs the centrifugal. If the curved isobars inclose the higher pressure (anticyclone), then the gradient force G and centrifugal force F act in the same direction, and the negative sign is used before $\frac{c^2}{r}$ in the above equation. If they inclose the lower pressure (cyclone) then G and F are opposed, and the positive sign is used. Figure 85 gives the scheme of the balance of forces for cyclonic and anticyclonic circulation in the northern and southern hemispheres. Since both forces balancing G always act at right angles to c, it remains true for curvilinear as for straight line motion that all stationary air currents must follow the isobars. Winds fulfilling these conditions are called gradient winds. From the above equation the gradient wind velocity for cyclonic and anticyclonic circulation may be computed in terms of the pressure gradient, radius of curvature, and latitude. Since in anticyclones the pressure gradient is balanced only by the difference between the deflective and centrifugal forces, the pressure gradient can never become as steep as in cyclones, and since of the above three forces it is only the pressure gradient which can increase the wind velocity c, the winds are less in anticyclones than in cyclones. (See Figures 71 to 74, Chapter IV, p. 121 and 122.)

The gradient wind velocity for cyclonic circulation is found from the solution of $\frac{1}{\rho}\frac{\partial p}{\partial n}=lc+\frac{c^2}{r}$ to be

$$c=\sqrt{\frac{r}{\rho}\frac{dp}{dn}+\left(\frac{rl}{2}\right)^2}-\frac{rl}{2}.$$

For anticyclonic circulation from $\frac{1}{\rho}\frac{\partial p}{\partial n}=lc-\frac{c^2}{r}$ it follows that

$$c=\frac{rl}{2}-\sqrt{\left(\frac{rl}{2}\right)^2-\frac{r}{\rho}\frac{dp}{dn}}.$$

Numerical values of c may readily be computed for any particular set of values of r, l, and $\frac{dp}{dn}$. Humphreys [10] gives a gradient velocity nomogram

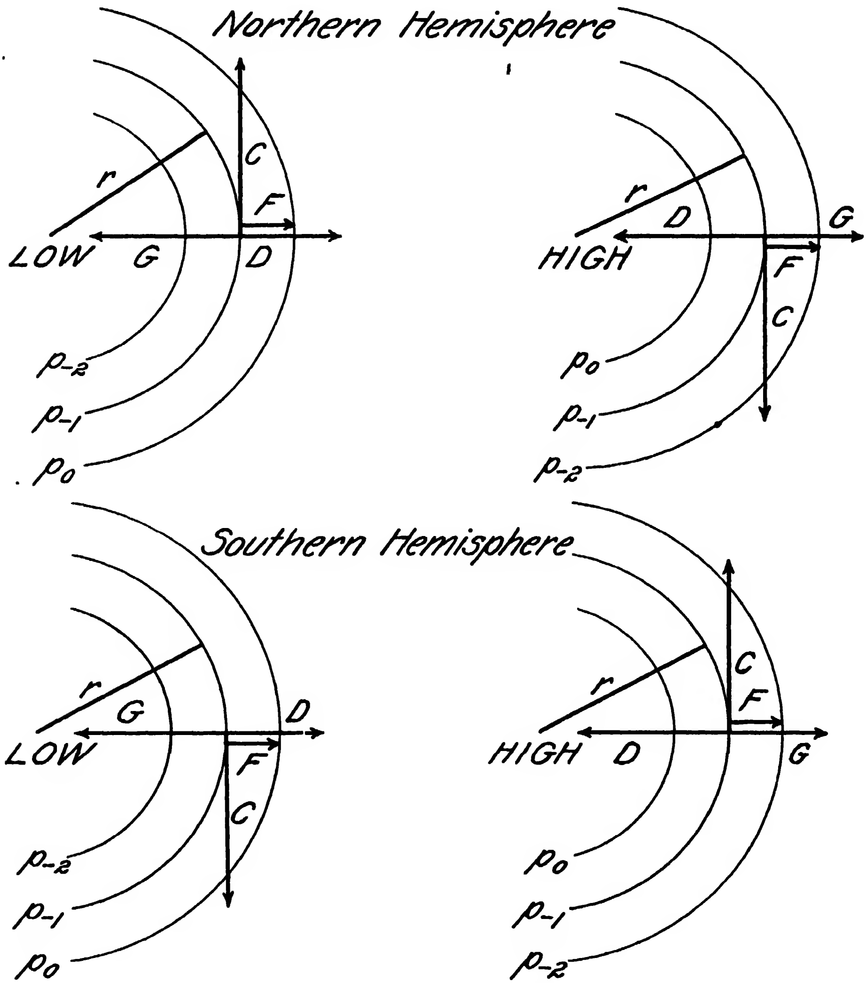

FIG. 85.—Gradient wind balance of forces for curvilinear air flow on the Northern and on the Southern Hemisphere.

or graphical solution of these equations, by means of which the numerical value of c may be quite accurately determined without any computation, if the latitude ψ, the radius of curvature r in kms., and the horizontal pressure gradient $\frac{dp}{dn}$ in mm. of Hg. per 100 km. are known. For surface

winds ρ may be regarded as constant. It is obvious from the above values of c that in cyclonic circulation there is no limit to the possible value of the pressure gradient $\frac{dp}{dn}$ for which there may exist a gradient velocity c such that stationary conditions can obtain. But in the anticyclonic circulation the limiting value of $\frac{dp}{dn}$ is given by $\frac{r}{\rho}\frac{dp}{dn}=\left(\frac{rl}{2}\right)^2$, and the corresponding maximum wind velocity by $c=\frac{rl}{2}$. This is the quantitative expression of the already noted effect on gradient wind velocity of the fact that in cyclones the pressure gradient is balanced by the sum of the deflective and centrifugal forces, in anticyclones by the difference. This maximum velocity is that which the earth's surface has at the same point due to the component of its rotation about the axis passing through the center from which the radius of curvature r is taken.

Friction.—Actual observations of surface winds show a consistent departure from gradient wind equilibrium. Synoptic charts show that the prevailing surface wind direction is not parallel to the isobars, but at an angle across the isobars from higher to lower pressure, and that this angle is greater over rough land areas than over the sea. Hence it must be concluded that at the earth's surface there is some retarding force at work which prevents the attainment of the full gradient wind velocity, and consequently prevents the balancing of the gradient force by the deflective and centrifugal forces. This retarding force is that of friction. Observations show that at an elevation of only half a km. the wind velocities approach quite closely the gradient velocities indicated by the synoptic charts. This indicates that friction at the ground surface is principally effective in slowing air movements.

Skin friction.—The first real attempt to take the frictional force into account in the equations of motion was made by Guldberg and Mohn.[20] They assumed that the effect of friction is limited to the air layer just above the ground due to the drag of the ground. This type of frictional force limited to a very thin surface or " skin " layer of the atmosphere is called skin friction. Guldberg and Mohn further assumed that the frictional force is proportional to the wind velocity, and active in the opposite direction. Thus the frictional force R for a wind velocity c is given by $R=-kc$, where k is a proportionality factor called the coefficient of friction. If account is taken of the frictional force the equations of rectilinear motion become $\frac{du}{dt}-lv+ku=-\frac{1}{\rho}\frac{\partial p}{\partial x}$ and $\frac{dv}{dt}+lu+kv=-\frac{1}{\rho}\frac{\partial p}{\partial y}$. For stationary motion, or $\frac{du}{dt}=\frac{dv}{dt}=0$, by squaring and adding the above

equations, the relationship $c\sqrt{l^2+k^2} = -\frac{1}{\rho}\frac{\partial p}{\partial n}$ is obtained. Furthermore, if Ψ is the angle between the wind direction and the pressure gradient for stationary motion, it is readily found that $\tan\Psi = \frac{l}{k}$, and that the limiting case of $k=0$ and $\Psi=0$ is that of the gradient wind. This gives two independent methods of determining k with the help of synoptic charts on which $\frac{\partial p}{\partial n}$, c and Ψ can in cases of practically stationary rectilinear air movement actually be measured, and thus the correctness of the theory can be closely checked. It was found that if values of k, found in this way from observed values of $\tan\Psi = \frac{l}{k}$, were used in the expression $c\sqrt{l^2+k^2} = -\frac{1}{\rho}\frac{\partial p}{\partial n}$, the calculated values of c thus obtained would be about double those actually observed. Similarly, if k is obtained observationally from the second equation, it gives theoretical values of Ψ which do not correspond at all satisfactorily to the observed values. Thus Guldberg and Mohn's assumption cannot explain quantitatively the observed relationships between wind direction, velocity, and angle of deflection for stationary rectilinear motion. Needless to say it is no more successful with curvilinear motion. Sandström [21] showed that for cases of practically stationary motion with straight isobars on the synoptic charts, the three forces, gradient, deflective, and frictional, are not nearly in equilibrium if the last named is calculated from observed values of Ψ by Guldberg and Mohn's method. In every case Sandström found a net residual force sufficient to effect considerable acceleration of the wind velocity. Hesselberg and Sverdrup [22] simply took the resultant of Guldberg and Mohn's frictional force and Sandström's residual force, and arbitrarily called this the frictional force, acting no longer in direct opposition to the velocity c but at an angle β to the right of the opposition direction. The angle β has normally a value of about 38°. The balance of forces for stationary motion with straight isobars is represented schematically in Figure 86. The relation is the same for curvilinear motion except for the addition of the centrifugal force. This method of treating the friction problem followed by Hesselberg and Sverdrup gives equations which fit the observed facts for surface air movements, but their assumption does not of itself explain physically the nature of the frictional resistance assumed. To do this it is necessary to take into account the internal friction or friction between air masses of different velocities.

(b) *Internal friction.*—If it were assumed that only skin friction acted on an air current moving over the earth's surface, there should exist a very thin layer of air slowed by friction at the ground, and a very rapid transition to full gradient velocities just above. If the internal friction of air were only that measured in the laboratory, the molecular friction due to the interaction of molecules between air currents of different velocity, this transition layer would be only a few centimeters thick in the vertical. The fact that the transition takes place gradually through some hundreds of meters shows that there is another frictional force or drag between air layers that works on a much larger scale than the molecular friction. This is nothing other than turbulence, or the

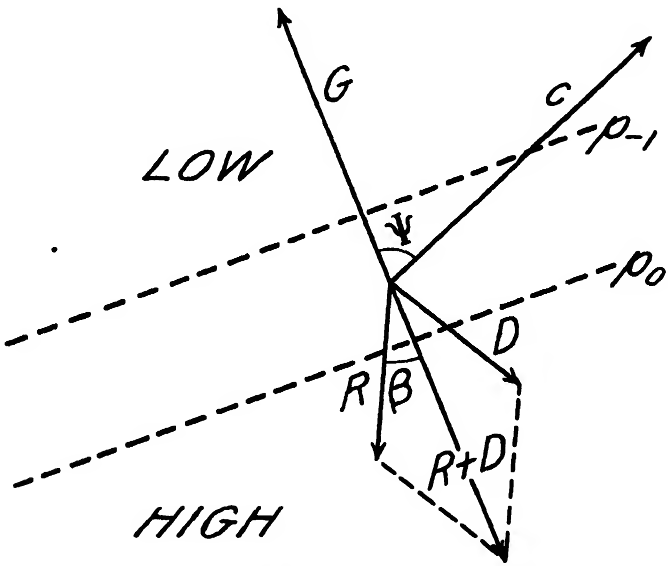

FIG. 86.—Friction according to Hesselberg and Sverdrup.

disordered interchange of considerable air masses instead of individual molecules between air layers of different momentum.

Under the discussion of small scale turbulence in the vertical direction it was found that the time rate of change of the property s per unit volume is given by $\frac{\partial s}{\partial t} = A \frac{\partial^2 s}{\partial z^2}$, where A is the austausch coefficient. The total change of s per unit volume due to the small scale vertical and horizontal austausch A (the turbulence austausch, to be kept distinct from the much greater horizontal austausch of the interzonal circulation of middle latitudes) is given by

$$\frac{\partial s}{\partial t} = A\left(\frac{\partial^2 s}{\partial z^2} + \frac{\partial^2 s}{\partial x^2} + \frac{\partial^2 s}{\partial y^2}\right),$$

and the change of s per unit mass by

$$\frac{\partial s}{\partial t}=\frac{A}{\rho}\left(\frac{\partial^2 s}{\partial z^2}+\frac{\partial^2 s}{\partial x^2}+\frac{\partial^2 s}{\partial y^2}\right).$$

If s be considered as momentum, then this expression gives the time rate of the change of momentum per unit mass of atmosphere due to the turbulent drag (momentum austausch) with the surrounding air masses of different momentum s. For unit mass the momentum s in the x and y directions is equal to the velocity components u and v, and the above expression gives the horizontal accelerations $\frac{\partial u}{\partial t}$ and $\frac{\partial v}{\partial t}$ per unit mass of air, resulting from the frictional drag of surrounding air masses, as

$$\frac{\partial u}{\partial t}=\frac{A}{\rho}\left(\frac{\partial^2 u}{\partial x^2}+\frac{\partial^2 u}{\partial y^2}+\frac{\partial^2 u}{\partial z^2}\right)\quad\text{and}\quad\frac{\partial v}{\partial t}=\frac{A}{\rho}\left(\frac{\partial^2 v}{\partial x^2}+\frac{\partial^2 v}{\partial y^2}+\frac{\partial^2 v}{\partial z^2}\right).$$

These are the well-known expressions of hydrodynamics for the frictional force, the austausch coefficient A being identical with the better known coefficient of friction μ. The simple equations of rectilinear motion as given on p. 178 may now be extended to take into account the frictional force, or turbulent drag. The horizontal deflective force and pressure gradient force will effect accelerations $\frac{du}{dt}$ and $\frac{dv}{dt}$ diminished by the amount of the corresponding frictional drag, or

$$\frac{du}{dt}-\frac{\mu}{\rho}\left(\frac{\partial^2 u}{\partial x^2}+\frac{\partial^2 u}{\partial y^2}+\frac{\partial^2 u}{\partial z^2}\right)=lv-\frac{1}{\rho}\frac{\partial p}{\partial x}$$

and

$$\frac{dv}{dt}-\frac{\mu}{\rho}\left(\frac{\partial^2 v}{\partial x^2}+\frac{\partial^2 v}{\partial y^2}+\frac{\partial^2 v}{\partial z^2}\right)=-lu-\frac{1}{\rho}\frac{\partial p}{\partial y}.$$

Since the horizontal variations of u and v are negligibly small compared with the vertical, the equations of rectilinear steady motion are

$$-lv-\frac{\mu}{\rho}\frac{\partial^2 u}{\partial z^2}=-\frac{1}{\rho}\frac{\partial p}{\partial x}\quad\text{and}\quad lu-\frac{\mu}{\rho}\frac{\partial^2 v}{\partial z^2}=-\frac{1}{\rho}\frac{\partial p}{\partial y}.$$

If it be assumed that the pressure gradient is constant with elevation so that the total wind velocity $c=\sqrt{u^2+v^2}$ approaches with elevation the gradient velocity c_0, that the axes of reference are chosen so that the total pressure gradient is in the direction of x positive and that u and v at the surface of the ground are zero, then the solution of these equations leads to the value of the total horizontal velocity at elevation z as given by $c=c_0\sqrt{1-2e^{-az}\cos az+e^{-2az}}$ where C_0 is the gradient velocity corre-

sponding to the constant value of $\frac{\partial p}{\partial x}$, and

$$a=\sqrt{\frac{\rho l}{2\mu}}=\sqrt{\frac{\rho\omega\sin\psi}{\mu}}.$$

The angle Ψ between wind direction and pressure gradient at the elevation z is given by $\tan\Psi=\frac{v}{u}=\frac{1-e^{-az}\cos az}{e^{-az}\sin az}$.

In order to fix by these equations the distribution of wind velocity c and the angle Ψ between the wind velocity vector and the constant pressure gradient vector, at different elevations, it is necessary only to fix the value a, which in turn depends upon the austausch or internal friction coefficient μ. The other quantities in which a is expressed are definitely known. The determination of μ has been made experimentally from

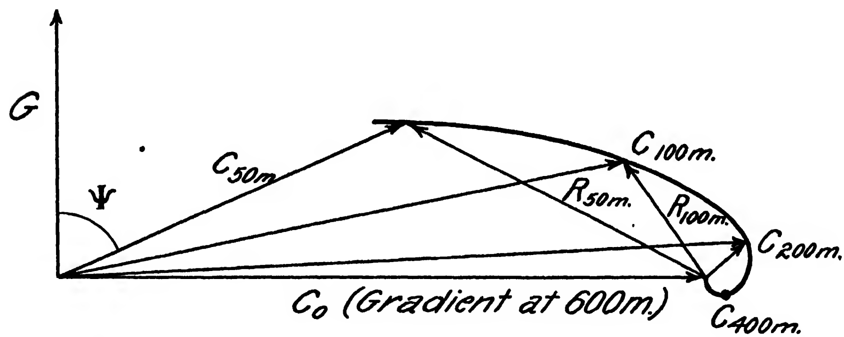

FIG. 87.—Variation of wind velocity and frictional drag with elevation.

series of observations of wind velocities on the Eiffel tower (Åkerblom), by upper air soundings above Lindenberg (Hesselberg and Sverdrup [22]) and by numerous others. The results obtained vary rather widely, but the best mean value found by Hesselberg and Sverdrup is that given by Schmidt of about 50 g.$^{-1}$/cm./sec.$^{-1}$ Hesselberg and Sverdrup point out, as Richardson has done, that the value of μ at a few hundred meters above the ground is many times what it is near the ground. Hence the variation of wind velocity and direction with elevation is much more rapid near the ground than higher up. The calculated distribution of wind velocity and angle Ψ based on the observed mean values of μ on the assumption of a constant pressure gradient sufficient to give a gradient wind of 20 m./sec., are indicated in Figure 87, and the accompanying table, taken from p. 121 of Exner's Dynamische Meteorologie. These calculations are based on Hesselberg and Sverdrup's results. The vectors C represent the velocities at the indicated elevations relative to the direction

of the pressure gradient G and gradient wind, the vectors R the corresponding frictional accelerations (net drag). The subscripts refer to elevations.

COMPUTED VARIATION OF WIND VELOCITY AND DIRECTION WITH ELEVATION

z Meters elevation	c Meters per second	Ψ Degrees
5	1.5	46½
10	2.9	48
20	5.7	48
50	11.7	59
100	17.9	70½
200	21.4	85
400	20.1	91
600	20.0	90
1,000	20.0	90

It will be noted that the ends of the velocity vectors lie on a logarithmic spiral, known as Ekman's spiral, because this theory was first developed by Ekman in the study of the effect of wind friction on ocean currents at different depths.

In general it may be said that the results obtained from the theory of the effect of skin friction at the ground and internal friction between the air layers above correspond very well with the actually observed distribution of wind velocity relative to the gradient velocity above the ground. Actually, conditions in the atmosphere seldom correspond to all the assumptions on which the above ideal velocity distribution in the vertical is based. In particular Hesselberg and Sverdrup have shown that normally the horizontal pressure gradient is far from constant with increasing elevation. This means that there is no constant gradient velocity c_0 aloft, but that internal friction is at work more or less between different air layers at all elevations.

The practical importance of the frictional forces in atmospheric circulations, general and local, cannot be overemphasized. In the discussion of the circulation principle it was pointed out that it is only through frictional resistance that the maintenance of a steady circulation in the atmosphere is possible. Another equally important aspect of the effect of the frictional force is that it permits of a considerable flow of air across the isobars from regions of higher to those of lower pressure. Under the gradient velocity equilibrium all air movement must take place parallel to the isobars, and no filling up of low pressure regions from high pressure regions can take place, once this equilibrium velocity is reached. The retarding action of friction, especially near the ground, has the

effect of making the ground act as a dissipator of horizontal pressure differences, so that the steepest horizontal gradients and greatest local variation of pressure are found in the free atmosphere aloft, in spite of the greatly reduced atmospheric pressure at higher levels.

Steady air currents and discontinuity surfaces.—Much of the discussion of the dynamics of air currents has been limited to the simplified conditions of steady currents. This is justified by the fact, repeatedly observed on synoptic charts, that large-scale air movements often tend to approximate the conditions of steady flow. No conditions of air flow in the atmosphere can be entirely steady; the frictional forces prevent it. However, the greater the extent of the air movements under consideration, the more insignificant, relatively, becomes the influence of friction. For this reason the great northward and southward moving air currents, frequently observed side by side in middle and northerly latitudes, may be considered practically as frictionless, and frequently as almost steady rectilinear motion. Helmholtz first, and later Margules, showed that air currents of different velocities and densities can remain in stationary equilibrium side by side only under the condition that certain fixed relations exist between the properties of the two air masses and the position of the transition zone between them. For mathematical treatment such a transition zone is always assumed to be a surface of discontinuity, and such discontinuity surfaces are considered of the greatest importance by the modern meteorologists. Two air masses of different temperature or density can be in equilibrium at rest only when the colder mass is spread out under the warmer, so that the discontinuity surface between them is horizontal. If the discontinuity surface is vertical no relative motion of the two masses can maintain equilibrium. If the colder mass lies in a wedge shape under the warmer, then the slope of the discontinuity surface must be such that the horizontal pressure gradient from the colder to the warmer mass is just balanced by the inequality of the horizontal deflective force of the earth's rotation acting on the two currents of different velocity. The best general derivation of the conditions of equilibrium of a discontinuity surface between two distinct air masses is that of V. Bjerknes,[23] whose principal argument is here very briefly sketched.

In the first place, it is evident that two boundary conditions must be fulfilled at the stationary surface between the air masses, the kinematic condition that there shall be no component of velocity normal to the boundary surface at the surface itself, and the dynamic condition that the pressure immediately on either side of the surface shall be the same. At the discontinuity surface then $p_1=p_2$, and $dp_1-dp_2=0$, where the

subscripts refer to the two separate air masses. From the equations of rectilinear motion in the frictionless form,

$$\frac{du}{dt}=\dot{u}=lv-\frac{1}{\rho}\frac{\partial p}{\partial x}\cdot\quad \dot{v}=-lu-\frac{1}{\rho}\frac{\partial p}{\partial y},$$

and $\dot{w}=-g-\frac{1}{\rho}\frac{\partial p}{\partial z}$, it follows, since for stationary conditions

$$dp=\frac{\partial p}{\partial x}dx+\frac{\partial p}{\partial y}dy+\frac{\partial p}{\partial z}dz.$$

that $dp=\rho[(lv-\dot{u})dx+(-lu-\dot{v})dy-(g+\dot{w})dz]$. For $dp=0$ this equation gives the slope of an isobaric surface. The relationship $dp_1=dp_2$ gives the equation determining the slope of the surface of discontinuity as follows:

$$[(\rho_1lv_1-\rho_2lv_2)-(\rho_1\dot{u}_1-\rho_2\dot{u}_2)]dx+[(\rho_2lu_2-\rho_1lu_1) - (\rho_1\dot{v}_1-\rho_2\dot{v}_2)]dy-[(\rho_1g-\rho_2g)+(\rho_1\dot{w}_1-\rho_2\dot{w}_2)]dz=0.$$

If for the sake of simplicity the line element of which dx, dy, and dz are the components is so chosen that dy vanishes, then the slope of the line element is given by $\frac{dz}{dx}$, and the slope of the isobaric surfaces in air masses 1 and 2 and the slope of the discontinuity surface between the air masses are given respectively by:

$$\frac{dz}{dx}=\tan\alpha_1=+\frac{lv_1-\dot{u}_1}{g+\dot{w}_1}$$

$$\frac{dz}{dx}=\tan\alpha_2=+\frac{lv_2-\dot{u}_2}{g+\dot{w}_2}$$

and

$$\frac{dz}{dx}=\tan\beta=+\frac{(\rho_1lv_1-\rho_2lv_2)-(\rho_1\dot{u}_1-\rho_2\dot{u}_2)}{(\rho_1g-\rho_2g)+(\rho_1\dot{w}_1-\rho_2\dot{w}_2)}.$$

Since in the case of two air currents moving without acceleration parallel to the y axis (gradient winds) $\dot{u}_1=\dot{u}_2=\dot{w}_1=\dot{w}_2=0$, for this case

$$\tan\alpha_1=+\frac{lv_1}{g},\quad \tan\alpha_2=+\frac{lv_2}{g}$$

and

$$\tan\beta=+\frac{l}{g}\frac{(\rho_1v_1-\rho_2v_2)}{(\rho_1-\rho_2)}.$$

Since $p_1=p_2$, $\frac{\rho_1}{\rho_2}=\frac{T_2}{T_1}$, and it follows that

$$\tan\beta=+\frac{l}{g}\frac{(T_2v_1-T_1v_2)}{(T_2-T_1)}.$$

Thus the slope of the discontinuity surface, for steady flow, must be greater the nearer together the temperatures of the two air currents and the greater the velocity difference. If subscript 1 refers to a colder southward moving current for which v_1 is negative, and subscript 2 to a warmer northward current for which v_2 is positive, then evidently tan α_1 and tan β are negative, indicating that the corresponding surfaces slope upward from right to left, while tan α_2 is positive, showing that the isobaric surfaces in the warm air slope upward from left to right (see Figure 88). Obviously at the discontinuity surface there is an abrupt angle in the slope of the isobaric surfaces, which will in this case of north

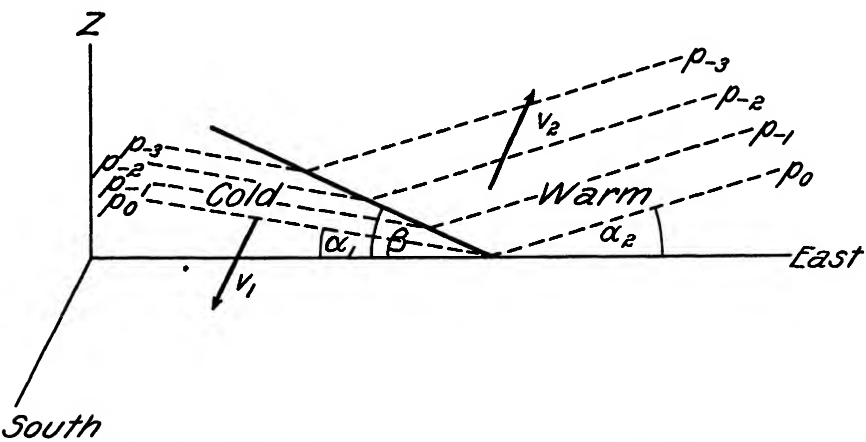

FIG. 88.—Discontinuity surface and isobars between oppositely directed warm and cold air currents.

wind on the west and south wind on the east, slope upward in either direction. Hence the minimum of pressure occurs at the discontinuity surface. This situation is represented in Figure 88, where the very small slopes of the isobaric and discontinuity surfaces which normally occur in the atmosphere have been exaggerated. Actually these angles are always very small, probably an extreme value of the discontinuity slope being 1/50, for the isobaric surfaces still less. In general the results of the above equations can be summed up by saying that two air currents of different density can remain stationary side by side only if the colder mass lies as a wedge under the warmer and flows to the right relative to an observer looking from the colder to the warmer mass. In the southern hemisphere the relative direction of flow is reversed.

From the fact that the colder air must always underlie the warmer in the form of a wedge, it follows that a warm air current imbedded in colder air must show a V-shaped cross-section whereas a cold air cur-

rent imbedded in warmer air must show an inverted V-shape cross-section resting on the ground. Margules [24] has derived mathematically an expression for the relation between horizontal temperature gradient and velocity distribution in a horizontal straight line air movement if the conditions are to be stationary. His results are summed up in the following two practical rules:

1) If the air aloft moves faster than the air at the ground, then to the observer looking in the direction of the upper air flow the higher temperature must be to the right, the lower to the left.

2) If the air aloft moves slower than the air at the ground, to the observer looking against the wind the higher temperature will be on the right, the lower on the left. In the first case the temperature and pressure gradients are directed together, in the second case they are opposed.

V. Bjerknes [23] has considered in detail the question of the effect of an unequal rotation of the atmosphere on the equilibrium position of a normally horizontal discontinuity surface between air masses of different potential temperature. He shows that if the warmer (upper) air mass is rotating more rapidly than the lower, a cyclonic sense of rotation tends to suck up the colder air from below and raise the level of the surface of discontinuity in the center of the vortex, and an anticyclonic sense of rotation tends to depress the level of the discontinuity surface. If the colder (lower) air mass rotates more rapidly, cyclonic rotation tends to suck down the warmer air, and hence the level of the discontinuity surface, while anticyclonic rotation tends to force up the discontinuity surface from below. The second case corresponds to the conditions existing in the atmosphere and the observed variation of the height of the base of the stratosphere above cyclones and anticyclones.

Besides the almost horizontal discontinuity surface at the base of the stratosphere, there are two other permanent discontinuity surfaces which are of importance in the general atmospheric circulation. These are both sloping surfaces between distinct wind systems. One is at some elevation above the ground between the trades and antitrades in the subtropics, the other is the so-called polar front between the westerlies of middle latitudes and the easterlies of the polar regions. These phenomena will be considered further in the treatment of the general atmospheric circulation. The variable temporary surfaces of discontinuity, stationary or non-stationary, horizontal or sloping, are of very frequent occurrence in the atmosphere. There are normally in the troposphere one or more almost horizontal discontinuity surfaces present, usually indicated by a temperature inversion and wind discontinuity, as shown by upper air soundings. They are due probably in part to the horizontal air currents in a thermally stratified atmosphere, in part to evapora-

tion of cloud layers, in part to unequal radiational cooling of different air strata, and in part to other causes. Their principal effect is to act as shielding layers (sperrschichte) against vertical convection. This effect is frequently visible in the structure of towering cumulus clouds, which tend to spread out beneath an inversion until at some point the upward movement acquires sufficient force to break through the inversion and tower up anew. For the sloping temporary discontinuity surfaces between moving air masses of different properties it was found above that very definite conditions must be fulfilled if they are to remain in stationary equilibrium. As a matter of fact, such surfaces are as likely to be in a non-stationary condition as not, and since it is the non-stationary condition which is of real significance for the state of weather, a careful consideration of non-stationary discontinuity surfaces is necessary.

Non-stationary discontinuity surfaces.—As long as a discontinuity surface fulfills the stationary conditions its only practical importance lies in its being the boundary surface between two air masses of different properties, and in the possibility that the conditions may cease to be stationary. As long as the slope of the surface is the equilibrium slope, there will be no vertical air movements. But as soon as the slope of the surface ceases to be the equilibrium slope, vertical accelerations along the slope must occur, which in turn are likely to lead to condensation and precipitation. The sense of the accelerations which occur is always such as to tend to bring the slope of the surface into accord with the equilibrium slope. If the extent of the surface were infinite, these accelerations would effect no change in the slope, but the reaction to these accelerations would be the force required to maintain equilibrium at the given slope. However, because these surfaces are not infinite, but are bounded below by the earth and above by the upper limits of the air masses they are separating, the vertical accelerations tend gradually to change the slope towards the equilibrium value. Thus, if the slope is too steep, the colder air must spread out under the warmer below, the warmer over the colder above, and *vice versa* if the slope is not steep enough. In these two cases the component of vertical velocity parallel to the discontinuity surface on each side of the surface is shown by the small arrows in Figure 89. It frequently happens, especially in the case illustrated by Figure 89a, that the vertical movement of the warm air upward is caused more by a general convergent wind flow and forced ascent of the warm air over the cold air than by any very strong tendency of the discontinuity surface to seek an equilibrium slope. However, usually the convergent flow in the first place is the result of a disturbance of the stationary flow of the air masses on either side of the discontinuity.

Non-stationary sloping discontinuity surfaces may be grouped in four classes according as the warmer (upper) air mass is 1) flowing actively upward along the surface, 2) moving passively upward due to an active underrunning and lifting up by the cold air beneath, 3) flowing actively downward along the surface, or 4) moving downward passively due to an active recession of the lower colder air from beneath. Classes 1) and 2), which are both characterized by warm air moving upwards over colder

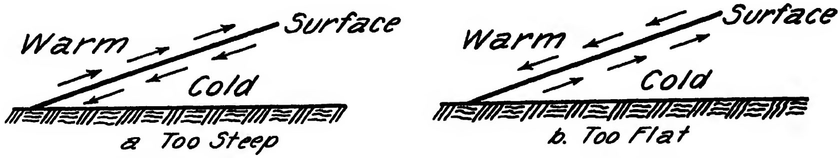

FIG. 89.—Vertical air movement at a surface of discontinuity.

air at the discontinuity surface, have been very appropriately named by G. Stüve [26] active and passive aufgleitfläche. They are the well-known warm and cold fronts of the Bjerknes cyclone model. Evidently the joint effect of a pair of such surfaces, both tending to a smaller slope, is to effect a spreading out of colder air under the warmer, and a decrease of potential energy of the mass distribution of the system with a corresponding increase of energy of motion, or kinetic energy. At this point

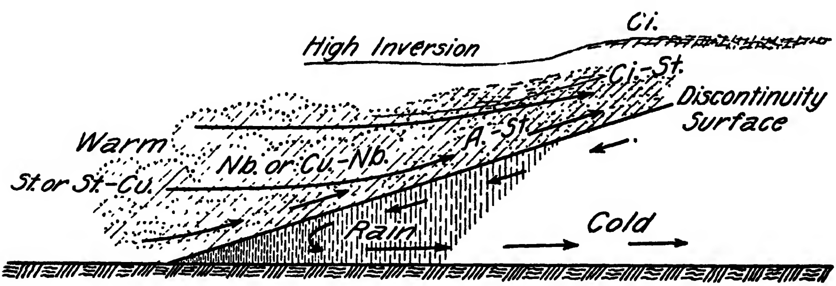

FIG. 90.—Vertical cross-section of a typical warm front.

these discontinuity phenomena can be described but briefly, as unrelated facts of observation. The important rôle they fill in the complex structure of cyclones and anticyclones must be left for discussion in connection with those phenomena.

The active aufgleitfläche, or warm front, normally has a slope of between 1/50 and 1/200. Because the ascending air current is warm and therefore normally of rather high specific humidity, the condensation phenomena characteristic of this type of front are very well marked. Figure 90 shows the typical warm front circulation and condensation

forms, after the scheme of Bjerknes [25] and Solberg. This gives a sloping cloud bank of decreasing elevation and increasing thickness as the cold air recedes and the warm front approaches. In one respect does this diagram depart from Bjerknes's scheme, namely, that of the individual Ci clouds which may be forerunners of the bad weather. Stüve [26] attributes these forms to a slight advance forcing up of an upper inversion with its high humidity at the under side, rather than to condensation in the front of the ascending air current. This latter shows as CiSt, quickly passing into ASt, *ASt praecipitans* (pallio-nimbus) and genuine Nb, or even Cu-Nb, if the thermal stratification in the warm air mass is only conditionally stable. The foremost portion of the rising air current has not, normally, come from the ground, but from a position in the warm air mass only slightly lower than that at which it is found and at some distance from the front. It is, however, the ground current in the warm air mass which is most lifted, and whose humidity determines the condensation level or height of the cloud base at the time of the passage of the warm front at the ground.

The cold front discontinuity surface is characteristically steeper than the warm front, perhaps normally from 1/40 to 1/80 in the foremost portion. In the case of the warm front the effect of surface friction at the ground is to retard the recession of the cold air, particularly in the hindmost portion where the cold air layer becomes very thin. The effect of this tendency to a frictional lag in the recession of the last portions of the cold air mass is obviously to decrease the slope of the warm front where it becomes close to the ground. The reverse effect occurs in the advance of the rapidly underrunning cold air at the cold front. The advance is retarded at the ground, so that there is a tendency to an overrunning by cold air aloft and a piling up of the cold air in the foremost portion of the advancing current. W. Schmidt [27] has shown by laboratory experiments with liquids that this squall head (böenkopf) really does exist, at least when a denser liquid underruns a lighter. The foremost portion of the invading cold air really rolls over in its advance, due to the retarding effect of friction at the ground. (See Fig. 91.) This accounts for the sudden rapid rise of pressure and squall which normally accompany the passage of the cold front. Probably there is frequently appreciable overrunning of the warm air in advance by the invading cold air just ahead of the passage of the cold front at the ground. This leads to the very marked convectional instability together with conflicting wind currents which are most favorable for the formation of tornados. However, it seems extremely unlikely that the cold air aloft can overrun the warm air beneath by hundreds of miles, as is frequently assumed in explaining instability showers far in advance of the front passage at the

ground. When upper wind directions seem to indicate this, it is much more probable that the more northerly winds aloft do not belong at all to the cold air mass whose front at the ground is several hundred miles to the west.

Abgleitfläche, or surfaces of subsidence (classes 3) and 4) above) are of less practical importance in weather phenomena than the aufgleitfläche, or warm and cold fronts. In general, their slope is very slight, perhaps 1/400 to 1/1,000, and correspondingly the air movements are very light. In fact, in the most frequent type of abgleitfläche, the surface of subsidence in the stagnant anticyclone, the lower colder air mass is completely at rest, the upper warmer air settling very slowly down the slope of the discontinuous surface, which is highest at the center of the anticyclone. In this case the discontinuity surface has been formed gradually in an originally cold homogeneous air mass of which the upper

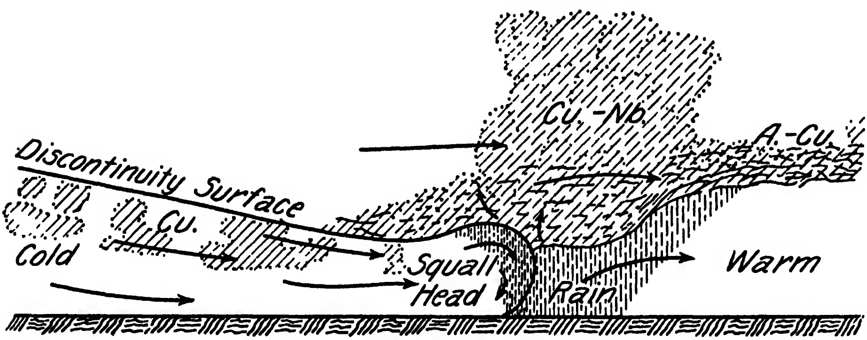

FIG. 91.—Vertical cross-section of a typical cold front.

portions have settled and spread out horizontally, as explained in the discussion of the effect of large-scale vertical air movements on the lapse rate. The discontinuity surface initiated purely by the dynamic heating of the upper air strata is quickly intensified by other factors. It becomes a definite boundary for the effect of mechanical turbulence from the ground. Although this turbulence is slight, such a situation may hold for weeks, especially in the interior of a continent in winter. Mechanical turbulence in the lower cold layer tends to effect uniform distribution of water vapor and potential temperature, and to carry dust and smoke upwards to the inversion. Thus the lowest temperature, highest relative humidity, and greatest concentration of dust and smoke occur at the base of the inversion, while at the top the air is warm, very clear, and very dry, not only in relative but also in absolute humidity. This makes such a surface a very effective surface for the radiation of heat to space from the lower atmosphere, as well as a surface of reflec-

tion of the direct solar radiation. The net result is that for weeks at a time in favorable locations, such as central Europe, there may prevail markedly anticyclonic conditions, the lowest kilometer, more or less, of the atmosphere being extremely cold, with almost constant low stratus or smoke and fog, while at an elevation between one and two kilometers there exists a temperature inversion of perhaps 10°, 15°, or even 20° C., a decrease of relative humidity from 98 to 100 per cent to as low as 5 or 10 per cent, and a change from dense fog to a brilliantly clear sky. Numerous instances of such inversions have been discussed by W. Georgii,[28] W. Peppler,[29] and many others. The explanation of the maintenance of the anticyclonic high pressure itself will be taken up in the discussion of anticyclones.

This completes the analysis of the general principles underlying all atmospheric movements. It remains to apply these principles to the general, or primary, circulation of the atmosphere and to secondary, or cyclonic and anticyclonic circulations.

VI. THE GENERAL CIRCULATION

In any discussion of the so-called general or planetary circulation, it must be remembered that any general scheme purporting to represent this circulation actually pictures only the mean or average condition obtaining over the globe. Probably at no one time has the actually existing atmospheric circulation in all its parts even closely approximated any ideal system yet devised. The disturbing influences at work are innumerable. Chief among these is the seasonal variation of the sun's declination, with the consequent shifting of the thermal equator and the tropical-polar temperature contrast on the northern and southern hemispheres. Even more disturbing is the irregular distribution of land and water surfaces, of mountain ranges and ocean currents. However, the unequal distribution of incident solar radiation over the face of the globe necessitates a compensating atmospheric circulation. The problem is to discover from a statistical analysis of the still very scanty observational data from all accessible regions in the atmosphere the general features of this circulation in the mean. If the picture thus obtained is found to meet the conditions placed by the existing distribution of the incoming and outgoing streams of radiational energy over the earth's surface, then it may be assumed to represent correctly the main features of the mean atmospheric circulation.

This discussion of the general circulation will be concerned primarily with only the Northern Hemisphere, for the observational data are much more numerous here, as a result of the more direct concern of the greater part of the civilized world in the conditions of this hemisphere. It might

be pointed out that all observations indicate an almost identical scheme of prevailing circulation in the Southern Hemisphere, except that there the general features are more clearly marked and intense, the irregularities less. This is a result of the less extensive and more symmetrically situated land areas in the Southern Hemisphere.

The general circulation on a uniformly frictional non-rotating earth.—The necessary and sufficient condition for the existence of a general circulation in the atmosphere is the unequal supply of heat in the form of insolational energy to the atmosphere in equatorial and polar regions. In the discussion of radiation equilibrium it was pointed out that solar radiation is practically the only source of the heat of the earth's atmosphere. It follows that the great excess of solar radiation received in tropical regions over that received in polar regions must tend to maintain a higher temperature in the earth's atmosphere in the tropics. This means a lifting of the isobaric and a sinking of the isosteric surfaces in equatorial regions and the reverse in polar regions. Thus numerous solenoids are formed running parallel to the latitude circles, and the tendency to a marked circulation between tropics and poles is established. According to the scheme of Figure 81, since the vector G will be inclined poleward and V equatorward, the general circulation must be poleward aloft and equatorward below. If the earth were not rotating, and if the distribution of land and water were uniform in such a way that no longitudinal pressure gradients were established, there would presumably exist one large direct circulation system between the equator and each pole. The air would slowly cool by radiation from the upper levels and sink over the circumpolar half of each hemisphere, and heat at the ground and ascend over the equatorial half. A steady circulation would be established. Frictional forces would tend to check this circulation, so that the work performed by the thermodynamic system in overcoming this resistance would require the consumption of a part of the excess heat received in the equatorial regions. The thermodynamic efficiency of such a system would be $\frac{Q-Q'}{Q}$ where Q is the excess of effective insolational over emitted terrestrial radiation in the equatorial half of the system, and Q' is the excess of emitted terrestrial over effective insolational radiation in the polar half of the system. Actually, however, the earth's rotation and the distribution of land and water effect a complete breaking down of the above single direct uniform circulation between equatorial and polar regions.

Effect of the earth's rotation on the general circulation.—In the discussion of the effect of the rotation of the earth about its axis on horizontal air movements (see page 175 *et seq.*), apart from the centrifugal effect

which was found to be negligibly small for most considerations, there were mentioned two influences that must be taken into account. These are the so-called horizontal deflective force, and the tendency to the conservation of angular momentum. The first of these is an effect which it is not difficult to consider quantitatively in the discussion of any air movement or steady circulation. It is a fictitious force F acting at right angles to the linear velocity v of the air current referred to the earth's surface, and is applied to account for the apparent deflection in the direction of the air current which deflection is in reality due to the rotation of the plane of reference beneath it. It is evident at once from the fact that this fictitious force acts always at right angles to the velocity v of the air current that it can never produce any linear acceleration of v, but merely a change of the direction of v. Hence steady flow of air from higher to lower pressure on the northern hemisphere will be deflected more and more to the right until an equilibrium velocity (gradient wind) is reached parallel to the isobars such that the pressure gradient force is just balanced by the horizontal deflecting force. When this state is reached, evidently all steady movement of air from regions of higher to regions of lower pressure must cease, except as the piling up of air masses in the equilibrium zone steepens the pressure gradient, or as the retarding effect of friction near the ground slows up the velocity of flow sufficiently to permit of a certain component of v directed across the isobars. Clearly the direct circulation between polar and tropical regions outlined above for a non-rotating earth is impossible for a rotating earth, for the equatorward moving branch of the circulation will rapidly be deflected to a westward moving current, and the poleward moving branch to an eastward current. Therefore, instead of a single direct circulation between equator and either pole, there are found several smaller zonal circulations, within each of which more or less uniform east-west or west-east velocities appear at the earth's surface, as outlined below. It is also evident directly from the equations of motion that no steady poleward or equatorward gradient flow of air can take place around the entire earth at any latitude, for that would require a westward or an eastward component, respectively, of the pressure gradient around the entire globe, which obviously is a physical impossibility. Regular circumterrestrial latitudinal air movement can take place only as a consequence of the prevention by friction of the attainment of full gradient wind velocity.

Before passing on to the details of the general circulation, brief consideration must be given to the second of the two influences mentioned above, namely the rôle played by the conservation of angular momentum in atmospheric movements. If, in the movements of bodies of air from latitude to latitude, angular momentum about the axis of rotation of the

earth were conserved, the zonal linear velocities relative to the earth's surface which this would entail in latitudinal displacements frequently observed are very great. The simple fact that no such velocities are ever observed proves conclusively that the principle of the conservation of angular momentum is not applicable to the movements of isolated bodies of air in the atmosphere, or even to large northward or southward moving air currents of limited longitudinal breadth. This is quite natural, for this principle holds only where the forces acting on the moving masses are entirely central. When isolated portions of the atmosphere are considered, this condition is not met at all, for besides the centrifugal force and the force of gravity, there are the frictional forces and even more the pressure forces exerted by the remaining atmosphere. The effect of the pressure forces is immediately to check the least tendency to excessive zonal velocity of any isolated air masses at any latitude, while the frictional drag at the ground tends at each latitude to impart to the atmosphere the same angular momentum as the earth's surface possesses; that is, to bring the air to rest relative to the earth. However, in the case of regular interzonal air movement of entire circumplanetary atmospheric rings (Helmholtz rings), zonal acceleration of a complete ring is possible without the resistance of an opposing pressure force. It is just possible, then, in the most regular (subtropical) branch of the general circulation that the conservation of angular momentum may play a part in effecting the observed general accelerations parallel to the latitude circles. The net result as indicated by the nature of the general circulation is the same when steady conditions prevail, whether it be the deflective force or the tendency to constant angular momentum which acts. The latitudinal pressure gradient is approximately balanced by the horizontal deflective force, but not completely because of friction. Whether the zonal velocities are to be attributed entirely to pressure gradient acceleration and subsequent deflection, or in part to the conservation of the angular momentum of circumplanetary air rings which are displaced latitudinally by the prevailing pressure gradient, the practical result is the same. However, since the existence of such a degree of regularity in the general circulation as to justify the assumption of circumplanetary rings of uniform movement seems extremely doubtful, and since the general features of the circulation are the same in any case, only the horizontal deflective force and the pressure gradient force are taken into account in the following qualitative explanation of the general circulation.

Qualitative explanation of the general circulation.—(a) *Circulation of the subtropics* (*trades-antitrades*).—The maximum effect of insolational heating is felt at the heat equator, which suffers a certain seasonal displacement that is of little consequence in a qualitative explanation of

the general circulation. This maximum of insolational heating at the equator effects a lifting of the isobaric surfaces and the establishment of a circulatory exchange of air with more northerly latitudes, in accordance with the circulation principle. However, owing to the horizontal deflective force of the earth's rotation, a deflection of the typical northward and southward moving air currents takes place, so that the southward moving current approaches the equator as a north, northeast, and eventually close to the equator, an east wind and the northward moving current as a south, southeast, and east wind. These are the well known trade winds. These winds become deeper as they approach the thermal equator, where they join as one great westward current of air extending to high elevations. At the Equator, however, they rise above the earth's surface, leaving the belt of calms at the surface known as the doldrums. Thus at the Equator the atmosphere as a whole is deficient in angular momentum relative to the earth's surface, or in other words, is rotating more slowly. It has been claimed that above the great equatorial westward current there has occasionally been observed an eastward current of high velocity. However, the regular existence of such circumplanetary current is quite inconceivable. It could be caused only by a strong west-east pressure gradient in the upper tropical atmosphere extending quite around the globe, which is of course a physical impossibility. There is no other way in which the upper air strata at the equator could acquire an excess of angular momentum over that of the earth's surface.

The upper branch of the subtropical circulation has its source in the upper levels of the strong westward current of air which has initially been fed in by the trades, at the same time that it was heating and rising. Under the influence of the pressure gradient the air aloft gradually flows out towards the poles, the current on the North Hemisphere being deflected as it moves poleward first towards the north and eventually towards the east. This current is known as the antitrades. Already at approximately 30° N. or S. this current appears as an eastward flow, which has cooled sufficiently to be sinking to the earth's surface in the high pressure belts at these latitudes. From these high pressure belts the descending air of equatorial origin may feed into the general west winds to the north, or into the trades to the south and work back towards the equator. This southward moving branch closes the path of the subtropical link in the general circulation between equatorial and polar regions. Only that part of the antitrades which continues polewards from the subtropical highs with the general westerlies of middle latitudes may take part in the complete circulation. The high pressure belts of the horse latitudes are essential to the closed link in the general circulation, which exists in the subtropics. Dynamically the existence of such a

closed link is necessitated by the deflecting force, which checks the northward movement of the equatorial outflow at about latitude 30°, and permits only a very slight northward progress of the air from that point. Consequently in these latitudes a piling up of air masses occurs, or a high pressure belt is formed, such that a gradient is established towards the equator at the ground, the gradient which is necessary for the existence of the trades. The calculations of A. Peppler,[30] based on numerous kite and sounding balloon ascents from May to September over the tropical and subtropical Atlantic Ocean, show that at only 4 kms. elevation the equatorward gradient disappears, and above that level the gradient is poleward from the equator to high latitudes. The subtropical highs themselves are, at the surface, regions of very light variable winds, or calms, but above 4 kms. where the poleward pressure gradient appears, there prevail westerly winds. Between the trades and antitrades there exists a transition layer, abrupt enough to be called a discontinuity surface. The thickness and elevation of this layer vary considerably with location and season. It has been studied only in very limited regions, primarily over the subtropical Atlantic, and notably by Sverdrup.[31] V. Bjerknes [23] has investigated the conditions under which a zonal discontinuity surface will be stationary. On the assumption that the atmosphere may be treated dynamically as a great circular vortex about the earth's axis of rotation he finds, as did Helmholtz with his circumplanetary air rings, that colder air must lie wedge-shaped under warmer air, and that the slope of the discontinuity surface must lie between the horizontal (the limiting case for isothermalcy between the bounding air masses) and the parallel to the axis of rotation of the earth (the limiting case for equal angular momentum or longitudinal velocity of the two bounding air masses at the surface). He finds that the observed slope of the trade-antitrade front and the polar front (see below) correspond to very probable discontinuities of temperature and velocity at these more or less permanent atmospheric discontinuities. However, as yet data are entirely insufficient either to confirm or to disprove the applicability of the general vortex theory in explaining the subtropical and polar fronts. Furthermore, most studies of the subtropical circulation have been limited rather much to the eastern Atlantic. It seems quite probable that further to the west average conditions are quite different, and that a similar difference holds between the eastern and the western subtropical Pacific. The fact that extensive intensifications of the subtropical high pressure belt occur over these two oceans results in the maintenance in the mean of a certain anticyclonic circulation. Consequently it is quite possible that over the western Atlantic and Pacific in the subtropics the winds on the average show a certain velocity com-

ponent from the south, certainly it appears quite frequently, anyhow. Thus the poleward movement of tropical air belonging properly to the antitrades may frequently be assisted considerably by the surface winds caused by the circulations belonging properly to these so-called "centers of action," the Azores High and the Pacific High. To whatever extent this occurs, the vertical trade-antitrade circulation becomes a horizontal anticyclonic circulation. (See Bergeron's circulation scheme, page 206.) Similar centers of action appear on the southern hemisphere. Obviously such complications render entirely impracticable any attempt at a general mathematical computation of the subtropical atmospheric discontinuity surface on the assumption of the existence of a circumplanetary circular vortex, or circumplanetary air rings.

(b) *The circulation of the polar caps.*—In the polar regions the prevailing circulation, as indicated by very scanty observational data, appears to be a thermodynamic direct partially closed circuit, as in the tropics. It is, however, less regular and less strongly developed than the trade-antitrade circulation. Especially is that portion of it nearest the poles almost unknown. Apparently there is a sinking of cold air masses over the ice-covered arctic, and this air moves southward under the influence of the prevailing gradient, being deflected gradually southwestward and westward. At about latitude 60°, or a trifle northward from there, this cold current reaches the southern limit of any regular progress. Here it may in part be warmed sufficiently to rise and return poleward aloft as the upper branch of the polar circulation, in response to the reversed gradient aloft, but in part it moves southward in irregular outbreaks of cold polar air at the ground, giving rise to the phenomena so well known as cold waves. Through the first 3 or 4 kilometers above the ground there is normally a rather marked and persistent discontinuity (the polar front) between the prevailing westerlies to the south and easterlies to the north, and correspondingly at the ground there must be a trough of minimum pressure between these two wind systems. At a few kilometers elevation this minimum disappears, and a poleward gradient with prevailing westerly winds prevails from the equator to the poles. Since there is always some slight movement of air masses across isobars from higher to lower pressure as a result of frictional forces, this entails a gradual normal northward flow of air from equatorial regions through the middle latitudes into the polar circulation. Thus air of southerly origin, under normal conditions, is continually feeding into the polar circulation aloft. Hence there is a gradual accumulation of cooling air masses over the arctic regions, which means a growing southward pressure gradient at low levels, the prevailing gradient mentioned above. Eventually there follows a pushing out of cold air at

some favorable point along the polar front, and a great outflow of excess polar air masses into the prevailing westerlies occurs. This effects a damming up of the normal air flow and establishes longitudinal pressure gradients counteracting the horizontal deflective force. This in turn permits the outflow of air from the arctic regions to move far southward and eventually to pass into the trade wind circulation.

(c) *Circulation in middle latitudes.*—From the subtropical belts of high pressure to the polar troughs of low pressure there is normally a pressure gradient at the ground which makes the intervening regions one of prevailing westerly winds. These are, therefore, practically speaking, gradient winds. Since this surface flow is just the reverse of that of a thermodynamically direct circulation between northerly and southerly latitudes, it may be said that the circulation of middle latitudes is a reverse circulation maintained by the two direct circulations bounding it to the north and to the south. In other words, it is a forced circulation, the force being applied by the subtropical belt of high pressure and the subpolar belt of low pressure which are maintained by the two direct links in the general circulation. From about 4 kilometers well up into the stratosphere a poleward gradient prevails from the equator to the poles. The frictional forces always acting to prevent the attainment of full gradient velocities effect a steady transfer of air from lower to higher latitudes in the regions of prevailing westerlies. The source of this air is to be found in the westerly currents of sinking air of equatorial origin in the subtropical high pressure belts. As already pointed out, part of these sinking masses feeds into the trades and returns equatorward. while part finds its way gradually northward in the general westerly winds of middle latitudes. At the polar front the gradual northward progress of these westerly currents takes place primarily in the upper levels, as the colder surface easterly winds are overrun by the warmer west winds above.

Throughout the zone of westerly winds the velocities probably approximate, in undisturbed conditions, the gradient velocities for the prevailing poleward gradient. Using Peppler's [30] data which apply to the months from May to September, Exner shows [32] that Peppler's observed wind distribution at different elevations over the Atlantic ocean from the subtropics to the middle latitudes agrees very well with the gradient velocities computed from the observed gradients. For the winter months both the prevailing winds and the latitudinal pressure gradients are greater than Peppler's summer values for the northern hemisphere. On the basis of his observed summer temperature distribution over the North Atlantic Ocean, Peppler computed from surface pressures and the hydrostatic equation the mean summer pressure distribution. His results, con-

tained in the following table, are important in that they represent probably the most accurate picture to be had of the mean pressure gradients giving rise to the general circulation. A similar calculation made by W. H. Dines [33] for the pressure above Canada, Europe, and the equator (Batavia) confirms Peppler's results very closely.

MEAN SUMMER PRESSURES IN MM. (PEPPLER)

Latitude	Surface	Kilometers elevation							
		2	4	6	8	10	12	14	20
10°	760	602	473	369	285	217	163	121	46
30°	766	606	473	367	281	212	158	116	45
50°	761	598	465	358	272	203	152	112	45

In particular it will be noted in Peppler's figures that the trade wind gradient passes over into the reverse gradient of the antitrades at 4 kms. elevation and that at about 20 kms. the poleward gradient, which reaches a maximum value at 10 kms., vanishes completely, presumably becoming a reverse gradient above. Dines' figures show the maximum poleward gradient at 8 kms., the reversal at 18 kms., and in general somewhat larger values of the poleward gradient. These differences are probably to be accounted for by the fact that Dines' observations are not entirely for the summer months.

The fact of the existence in middle latitudes of a poleward pressure gradient to the height of 18 kms. or more raises the question as to how the equatorward flow of air to compensate the poleward flow in the prevailing westerlies takes place. It was formerly assumed that a uniform return flow occurred at intermediate levels, perhaps at about 10 kms., above the westerlies. But the existence of a poleward gradient at these levels renders impossible the existence of any uniform return flow here. And although there is apparently a reversal of gradient above 18 or 20 kms. it is inconceivable that the return flow should take place in the upper tenth or less of the atmosphere. That would require very high wind velocities, whereas the indications are that above the base of the stratosphere (10 to 15 kms.) the winds decrease with elevation. The only remaining possibility is that the return flow of air to the equator in middle latitudes does not occur regularly, but irregularly. The explanation is to be found in the great irregular outbreaks of air from the arctic regions, in the manner described in the preceding section. When an excess of air has accumulated in the arctic, eventually a general equator-

ward movement of the polar air masses occurs at some point. The establishment of such an outflowing current from the polar regions invariably favors the establishment of a northward moving current of warm air on its east side, with the existence of a trough of minimum pressure between the two, the pressure gradient counteracting the deflective force. In such a trough between two oppositely directed air currents occur the interactions which constitute a family of cyclones (see next chapter). In general, it may be said that in middle latitudes the circulation or exchange of air masses between higher and lower latitudes takes place irregularly, in the alternate cold and warm currents which are associated with cyclones and anticyclones. The normal undisturbed condition is one of uniform slow poleward movement at all elevations up to 18 kms., which condition is repeatedly interrupted by outbreaks of polar air in great southward currents which compensate both the accompanying poleward current of warm air and the normal uniform poleward movement of middle latitudes. It is a well-recognized fact that at such times the cold current is normally both deeper and stronger than the warm current. Occasionally, when upper air data are obtained, it is found that the cold current attains very high velocities at great elevations. Instances of this have been found both over Europe and over the United States. J. Georgi [34] has recorded several instances of upper air data for Iceland for the summer months which showed extremely high velocities from the north in the upper troposphere in polar outbreaks which were comparatively light at the surface. He found in some instances, in spite of the decreasing density with elevation, that the maximum southward transport of air mass occurred at as high as 10 kms. North winds in excess of 100 miles an hour may be found at this level, but no such southerly air currents have been observed.

The poleward heat transport effected by the irregular circulation of the middle latitudes has been discussed under horizontal austausch (see page 167). Defant's calculation [15] of the horizontal austausch over northern Europe from a statistical treatment of wind observations at Potsdam agrees very well with the value which, according to R. Mügge,[14b] is necessary at these latitudes to effect the poleward heat transport requisite to maintain the observed stratosphere temperatures, on the assumption that they represent radiation equilibrium. The irregular distribution of land and water, especially on the northern hemisphere, is most favorable to the unordered circulation of middle latitudes. The great contrasts of temperature which normally exist over land and water surface, especially in winter, favor the establishment of longitudinal pressure gradients and thus make possible northward and southward moving air currents and the genesis of cyclones and anticyclones. (See Bergeron's circulation

scheme below, Figure 92.) Although in this manner the irregular distribution of land and water surfaces greatly favors the irregular circulation of middle latitudes, it is not the essential condition. Under the existing pressure distribution and scheme of general circulation over the earth's surface, the circulation of middle latitudes could not be other than irregular.

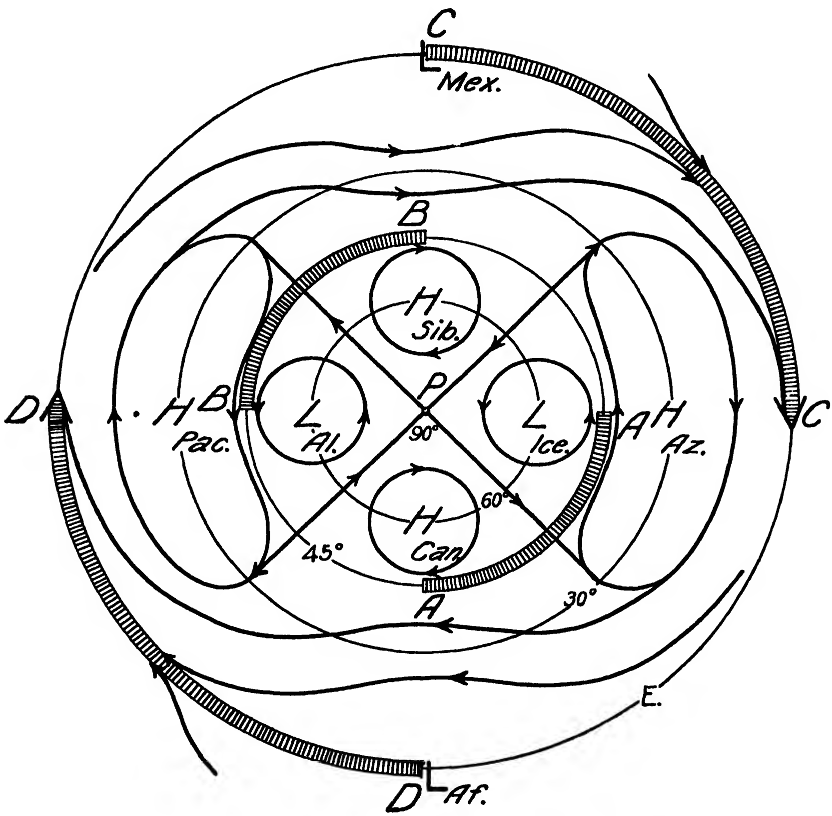

FIG. 92.—Bergeron's scheme of the general circulation of the Northern Hemisphere.

The way in which the actual distribution of land and water and consequent variations in surface temperature on the northern hemisphere work out in the mean is excellently shown by the above schematic representation taken from Bergeron[39] (Figure 92). This scheme represents the principal features of the normal air movements of the northern hemisphere in winter. The most important "centers of action," *i. e.*, semipermanent cyclones and anticyclones of such persistence that they are to be found on pressure charts representing mean conditions for the

season, are indicated by *L* and *H* respectively. These are the cyclones and anticyclones, best marked in winter, which form over warm ocean and cold continental regions (see next chapter). The continuous light lines indicate lines of flow, the direction being indicated by the arrows.

It will be noticed how the air movements take place around these centers of action as if they were two sets of interlocking wheels, and how the air currents of tropical or polar origin are brought from southerly to northerly or from northerly to southerly anticyclonic and cyclonic circulations. In this way the irregular distribution of land and water helps to effect the interchange of air masses between equatorial and polar regions. It will also be noticed that there are certain narrow zones, indicated by the heavy lines *AA* and *BB,* where currents of polar and tropical origin are brought together. In other words, these are regions where there is a tendency to the formation of very steep poleward temperature gradients and consequently atmospheric discontinuities, or fronts. To use Bergeron's term, they are regions of frontogenesis. Even the surface ocean currents tend to favor the sharpening of poleward temperature gradients in these two zones. To the north of the zone *AA* there flows in general from the northeast the cold Labrador current, on the opposite side from the southwest the warm Gulf stream, corresponding to and perhaps in part the consequence of the prevailing air flow. Similarly, to the north of the zone *BB* there flows from the northeast the cold ocean current from the Bering straits, and on the opposite side the warm Japanese current from the southwest. These regions are therefore the ones most favorable for the development of the migratory cyclones and anticyclones, and of rapid movement of the formations. The regions *AB* and *BA* on the contrary are regions of divergent air flow, hence of front destruction or to use Bergeron's term frontolysis. They are therefore regions of degenerating or filling cyclones, and of slow movement. This scheme agrees perfectly with the observed cyclonic and anticyclonic activity of these regions, as the discussion of the next chapter proves. The equatorial zones *DD* and *CC* represent regions of closely converging trade wind circulations, of well-marked doldrums, favorable for the development of tropical hurricanes. The regions *CD* and *DC* are characterized rather by weak divergence, or probably almost parallel flow of the two trade systems, therefore of prevailing east winds which are unfavorable to the genesis of tropical hurricanes.

In the Southern Hemisphere, on account of the much smaller ratio of land area to water area, and the central location of the only great land mass in high latitudes (the antarctic continent) the circulation in high latitudes is much more regular than in the Northern Hemisphere. All the features of the regular general circulation, the polar high pressure,

subpolar trough of low pressure, and subtropical belt of high pressure are more strongly developed and thus effect the same latitudinal exchange of air masses which is favored so much in the Northern Hemisphere by the irregularity of land and water distribution.

(d) *Circulation of the stratosphere.*—Both the observed values of atmospheric pressure at high levels and the values computed by the hydrostatic equation from the observed pressure distribution at the earth's surface and temperature distribution throughout the atmosphere indicate that the poleward pressure gradient vanishes at from 18 to 20 kms. Hence, presumably the prevailing westerlies and northward transport of air of the general circulation prevail up to about that elevation. Since the base of the stratosphere is known to be higher and colder over the tropics than over the polar regions, it follows that at about 12 kms. elevation the poleward temperature gradient of the troposphere becomes reversed. Thus interzonal air flow in the substratosphere effects a heat transport just the reverse to that of the troposphere. Southerly winds are relatively cold, northerly ones relatively warm. Since wind velocities reach a maximum at these elevations, temperature variations here are very great, and variations in the elevation of the base of the stratosphere are large. Because of the comparatively small density of the atmosphere above 12 kms. the actual equatorward heat transport effected in this upper branch of the general circulation is slight compared to the poleward transport in the troposphere; however, there are occasionally considerable pressure variations at the ground which can be explained only by the mass displacements effected by strong interzonal air currents in the substratosphere. This matter will be taken up again in the discussion of cyclones and anticyclones.

Above 18 or 20 kms. both the temperature gradient and the pressure gradient are the reverse of those in the troposphere. Therefore it seems reasonable to assume a secondary circulation in the upper stratosphere just the reverse of the general circulation of the troposphere and substratosphere. Whether this really exists or not cannot be determined from the scanty existing observational data from very high levels. In any case, this circulation is of no practical significance in the general thermodynamic equilibrium of the atmosphere, for in consequence of the very great elevations at which it occurs the total atmospheric mass and the horizontal pressure gradients must be so small that the horizontal mass and heat transport are negligible.

The control of the general circulation by thermodynamic and radiation equilibrium.—Certain fundamental facts of observation which have been stated at different points in this paper must be briefly recapitulated at this point and brought into harmony with one another. They are

four in number, namely 1) The base of the stratosphere varies in mean temperature from about $-80°$ C. over the equator to about $-45°$ C. over the poles, and its elevation from about 16 kms. over the equator to about 7 kms. over the poles. 2) The distribution of solar insolation is such that in the course of a year only 40 per cent as much heat is received per unit area at the poles as at the equator. Quite possibly the effect of the snow and ice cover over the polar caps reduces still further the ratio of effective insolation at the poles to that at the equator. 3) The intensity of the terrestrial radiation to space is appreciably greater at the poles than at the equator, as indicated by the higher temperature of the stratosphere at the poles, which represents radiation equilibrium practically speaking under the influence of terrestrial radiation only. That the earth can radiate more heat to space in the polar regions in spite of the lower surface temperatures there, is due to the relatively small amount of water vapor in the atmosphere, and hence the resulting low level of the effective radiating surface of the earth in these regions, and to the horizontal transport of heat from the Tropics by the general circulation. 4) The general circulation, in spite of its irregularity and breakup into separate parts by the earth's rotation, is in the mean, *in toto,* a thermodynamic engine acting between the polar cold source (net loss of heat to space from the upper atmosphere by radiational exchange) and the tropical heat source (net gain of heat from space to the lower atmosphere by radiational exchange). The work done by this engine consists in maintaining the circulation against frictional forces, and the circulation in turn effects the heat transport necessary for the continuance of such warm and cold sources.

Of these four facts, the temperature distribution in the atmosphere the distribution of effective insolational heating, the distribution of terrestrial radiation to space, and the general circulation, only one, the distribution of insolation, is definitely fixed. The other three quantities are mutually interdependent, and the mean prevailing conditions represent a very delicate adjustment or equilibrium condition, in which the water vapor in the atmosphere plays an important and complicating rôle. If there were no water vapor, *i. e.,* no element variable with temperature absorbing terrestrial radiation, equilibrium conditions in the atmosphere would be about as follows: 1) A radiation of heat to space at each latitude approximately proportional to surface temperatures (neglecting variations in the surface coefficient of radiation). 2) A poleward temperature gradient at all elevations much less than the present gradient in the lower troposphere. 3) A stratosphere somewhat warmer over the equator than over the poles, and only very slightly higher. 4) A

single rather weak general circulation system, involving both troposphere and stratosphere. This thermodynamic circulation would be weak because the temperature difference of the sources must be small, a small heat transport being sufficient to supply the difference between effective solar insolation and terrestrial radiation at each latitude. Equilibrium would be established between the temperature or density distribution, the latitudinal heat transport (to supply the radiational output-income difference), and the general circulation, which last will work against the frictional forces with a certain definite efficiency in effecting the necessary latitudinal heat transport.

The actual general atmospheric equilibrium may be explained qualitatively by the modifications of the above ideal case effected by water vapor. By selective absorption of the long wave terrestrial radiation in preference to the short wave solar radiation, the presence of water vapor in the troposphere effects a considerable warming of the lower levels of the atmosphere. This effect is much greater in the equatorial regions, where the higher temperatures make the presence of much greater quantities of water vapor possible than in the polar regions. Thus the presence of water vapor in the atmosphere effects a great increase of the poleward temperature gradient in the troposphere in the above ideal circulation scheme. This means an immediate increase of the interzonal circulation in the lower levels and the transport of much greater amounts of heat to the poles, hence an increase of the polar radiation to space, and a decrease of the equatorial radiation. Equilibrium is reached only when considerably more heat is radiated to space at the poles than at the equator. This in turn means a relatively warm stratosphere over the poles and a relatively cold over the equator. Since the difference between surface and stratosphere temperature is much greater at the equator than at the poles, and since the prevalence of convective equilibrium in the troposphere rather definitely fixes the vertical lapse rate, the stratosphere must be much higher in equatorial than in polar regions, as well as colder. The hydrostatic effect of the cold stratosphere, as pointed out under the discussion of the general circulation in the stratosphere, is to cause a reversal of the poleward pressure gradient at 18 or 20 kms. This reversal, in conjunction with the reversed temperature gradient, probably sets up a slight circulation counteracting to a small degree the thermal effects of the principal interzonal circulation. Such is the complicated nature of the equilibrium which prevails in the atmosphere between the radiation and the thermodynamic advection processes in the presence of water vapor.

VII. SECONDARY CIRCULATIONS—CYCLONES AND ANTICYCLONES

In the discussion of the general circulation it was pointed out that for interzonal exchange of air masses in middle latitudes the existence of longitudinal pressure gradients is a necessary condition. For this reason the stationary and migratory cyclones and anticylones of the temperate zones are of fundamental importance. They are nothing other than limited regions of locally deficient or excessive atmospheric pressure with a corresponding individual circulation. It is these secondary circulations, so important because they effect interzonal air transport, which are to be considered at the present point.

In the first place it must be realized that these cyclonic and anticyclonic circulations are extremely complicated phenomena. The interaction of mechanical and thermodynamic forces involved remains quite unexplained quantitatively, and very unsatisfactorily so even qualitatively. Any explanation of these most important meteorological phenomena is at best only roughly qualitative and in parts entirely hypothetical. The cyclone or the anticyclone must be considered from two distinct points of view; that of the pressure distribution and that of the kinetic energy of the accompanying circulation. Since vertical accelerations are negligibly small compared to the acceleration of gravity, the surface pressure distribution represents simply the weight of the overlying atmosphere, in accordance with the hydrostatic equation. The closed surface isobars characteristic of cyclones and anticyclones inclose regions above which the atmospheric air column is deficient or excessive in mass, respectively. According as this deficit or excess of mass is located near the ground or at higher levels, the formation has its maximum intensity at the ground or aloft.

Without the existence of the proper circulation, a well marked cyclone or anticyclone could not exist. In the discussion of gradient winds it was pointed out that apart from the effect of friction, stationary air flow can take place only parallel to the isobars, the velocity being such that the deflective force, centrifugal force, and gradient force balance each other. Thus only the circulation of a cyclone or anticyclone prevents the rapid equalization of pressures and the disappearance of horizontal gradients. Hence the center of high or low pressure can increase in intensity only as the circulation grows with it. The two must increase together, and similarly decrease together, at least to that degree in which cyclonic and anticyclonic circulations may be considered as representing steady conditions. The effect of friction being to keep the wind velocities, especially near the ground, at somewhat less than gradient velocities, it follows that the maintenance of a cyclonic or anticyclonic circula-

tion in undiminished vigor requires a constant supply of energy to make good this tendency towards an equalization of pressures near the ground. The growth of a cyclone or anticyclone requires a still greater supply of energy, not only to overcome friction but to increase the kinetic energy of the system. The source of the kinetic energy of one of these secondary circulations is not to be looked for in the horizontal pressure gradient of the circulation itself. Margules has shown in particular instances [35] that the kinetic energy in a typical cyclone is some twenty times the potential energy of the horizontal pressure gradient within the cyclone. Since the horizontal pressure gradient is the only accelerating force which has developed this circulation, this can only mean that there has been some continuous process at work supplying the energy necessary for the maintenance and growth of the horizontal pressure gradient together with the acceleration of the circulation. In the last anaylsis this is always effected by some thermodynamic process (unequal supply of heat), though the energy may appear in the immediately preceding form as potential energy of the vertical mass distribution or as kinetic energy of the general circulation. But whatever its source, it must have the effect of increasing the number of solenoids within the region of the secondary circulation.

Regions of locally excessive or deficient atmospheric pressure at the ground are probably produced by one or more of the following processes in the atmosphere: 1) Regional heating or cooling and consequent expansion or contraction of the overlying air column and outflow or inflow of mass. 2) Mechanical interaction between air currents of different movements and densities with the formation of waves or vortices at the boundary between the currents. 3) Advective transport of warmer or colder air masses and consequent change in the mass of the overlying air column. Each of these three processes requires careful consideration regarding its *modus operandi* and the possible sources of the kinetic energy of the resulting circulation.

Regional heating or cooling and the formation of stationary cyclones and anticyclones: tropical hurricanes.—Regional heating or cooling of the atmosphere at high levels may result from a net gain or loss of heat by radiation, or by condensation or evaporation. The temperature of the stratosphere, and consequently the height of its base, depend upon the amount of long-wave radiation received from the earth's surface below. It has been pointed out that water vapor in the lower atmosphere decreases the amount of radiation passing out to space, causing the cold high stratosphere of the tropics. R. Mügge [14] has tried to account for the frequently observed transformation from the low anticyclone (characterized by cold air near the ground) to the more permanent high anti-

cyclone (characterized by a high cold stratosphere), as a result of the rapid increase of the moisture content of a cold polar air current moving southward. Although this may be a contributing factor, no study has been made of the time rate at which such a change can take place. There is good reason to think that this change from the low to the high anticyclone type is much more closely tied up with the interzonal transport of air masses in the substratosphere than with radiation phenomena. This process will be considered in more detail during the discussion of the advective transport of warm and cold air masses as a cause of cyclones and anticyclones.

Cooling of air layers by evaporation is probably of no significance in the formation of anticyclones. It can never be comparable with the heating caused by condensation, because any considerable condensation of water vapor mostly falls out as precipitation, so that the process is not reversible. Only very locally, in the thunderstorm, where heavy precipitation falls a considerable distance through warm dry air, does evaporation cooling have any fundamental dynamic significance. The heat of condensation, on the other hand, is a factor not to be neglected in any cyclone. It contributes materially to the heating and displacement of mass with consequent reduction of pressure in the cyclone. However, calculations readily show that in temperate regions the energy liberated by the condensation of the normal amount of precipitation is small in comparison to the kinetic energy of the great cyclonic circulations frequently occurring. Thus for the extratropical cyclone the heat of condensation can be only a very secondary cause of growth or source of energy. In the tropics and subtropics, on the other hand, where temperatures and moisture content are invariably high the expansional cooling of rising air causes a much greater amount of condensation and so furnishes a much greater amount of energy than in more northerly latitudes. The result is the well-known tropical hurricane or typhoon, a very intense cyclone, peculiar in that practically its sole source of energy is heat of condensation. Since there is no corresponding reverse process, the tropical cyclone, unlike the extratropical cyclone, does not occur in conjunction with an anticyclone; there is no such thing as a corresponding tropical anticyclone.

Concerning the details of the growth and structure of the tropical hurricane, very little is known, due primarily to the impossibility of getting upper air data in the active region, and to the rapidity with which these hurricanes go to pieces on passing inland. The condition favorable to the formation of tropical hurricanes is one of light winds, great warmth and very high moisture content. They never form closer to the equator than six or seven degrees. Hence the regions most favorable for

their formation are in the doldrums at the time of the maximum displacement from the equator, which in the northern hemisphere is August and September, in the southern hemisphere February and March. They develop most frequently where the belt of calms is most marked, or over any warm tropical sea where the prevailing winds are light and especially in the belt of calms between two more or less marked conflicting wind systems. Such regions are found, for example, between the two trade systems in the region of the Cape Verde Islands, in the western Caribbean, and in the seas generally southeast of Asia. (See Bergeron's Circulation Scheme, Figure 92.) They are sometimes found also on the south coast of Asia and on the Indian Ocean between the monsoon and trade wind systems.

During the initial stages of the development of the tropical hurricane, before any unified centered cyclonic circulation has made itself apparent, the immediate vicinity of the developing storm is for from one to five days marked by low irregularly fluctuating pressure, damp oppressive heat, and frequent local squalls and convectional showers of torrential rain. Once the regular cyclonic circulation has become evident, the further development of the storm to hurricane intensity takes place very rapidly, if conditions are favorable. The fully developed hurricane is the most severe and intense type of cyclone, covering the least area and having the lowest pressure and strongest winds. Only the tornado exceeds it in intensity. In contrast to the extratropical cyclone, the distribution of temperature and precipitation about the center of the tropical storm is quite symmetrical. There is no warm and cold sector, no noticeable temperature discontinuities, no marked rain zones. The precipitation is nearly equally distributed on all sides of the center, and is excessively heavy. Furthermore, these intense tropical storms are characterized by a small central calm, or "eye" of the storm, a few miles in diameter, where the winds are light, the clouds high or even broken, and the pressure near its lowest. Probably the more intense extratropical cyclones have a similar central structure, but this is usually less clearly distinguishable. Finally, in contrast to the cyclones of higher latitudes which move generally from west to east with the prevailing winds of middle latitudes, the hurricanes move at first northwestward, evidently with the antitrades (southwestward on the Southern Hemisphere) and not southwestward with the surface trades. If they pass inland over a continent, they dissipate very rapidly, much more so than the extratropical storms. If they do not pass inland first, somewhere between 20° and 30° latitude they recurve poleward and then eastward, many of them eventually passing into the regular paths of the extratropical storms, and travelling great distances. After their recurvature they rapidly acquire

the characteristics of the extratropical cyclone. Mitchell, who has made a special study of the movements of the tropical hurricanes originating from the Caribbean Sea eastward, has found [86] that the recurvature of the tropical hurricanes as they come northward is determined by the large anticyclones to the north. A complete anticyclonic barrier to the north may entirely prevent recurvature of the tropical storms, which will then tend to move into and up through low pressure troughs. Apparently they are carried along by the wind systems of the large anticyclonic circulations, probably at some 3 or 4 kms. elevation, just as initially their movement was that of the antitrades aloft.

No serious attempt at even a qualitative explanation of the genesis, growth, and structure of the tropical cyclone has been made. The following are a few of the more evident possibilities of the process. In the first place, the initial stages in the development require light winds, warm moist air, and active local convection. The initial pressure reduction is doubtless due to these factors, especially the large scale liberation of heat of condensation in the heavy convectional showers. Strong winds would interfere with the development of any local area of reduced pressure in this way. The very gradual inflow of air to the region of locally reduced pressure is little by little deflected by the earth's deflective force, and by very slow degrees the cyclonic gradient wind circulation is established. The nearer the Equator the disturbance is located the more slowly does this occur. It is also frequently asserted that the oppositely directed wind currents on the opposite sides of the low pressure trough (doldrums) may be equally important in the establishment of the cyclonic rather than anticyclonic circulation, though quite frequently these opposing wind systems seem to be at too great a distance from the incipient cyclone to affect its development. Moreover, frequently tropical hurricanes develop, as in the Bay of Bengal, where there is no conflicting wind system, at least at the ground. At any rate, after the establishment of the gradient wind circulation around the center, there can be an air movement in towards the center only near the ground, where the effect of friction is felt, and where the air is warmest and moistest. Here it feeds into the general ascending currents, the moisture is condensed in great amounts, the pressure further reduced by the consequent heating and outflow above, and a powerful vortex is quickly established. Very soon such velocities are reached that because of the small radius of the vortex the centrifugal force outweighs the horizontal deflective force from ten to twenty times in compensating the pressure gradient in the gradient wind equilibrium. The winds blow in across the isobars normally at an angle of about 15° in the right-hand forward quadrant of the advancing cyclone and fall off to about 25° in the left-

hand rear quadrant. Assuming the usual approximation to gradient winds at 500 m. elevation, a simple calculation shows that the air mass flow into the rising vortical portion of a typical cyclone is just about that required to furnish the amounts of moisture necessary to supply the usual copious rainfall. This air flows out again at high elevations, as shown by the upper cloud movements.

Observations of hurricane wind velocities show [37] that in the storm proper, in to the outer limits of the central calm, the distribution of velocities is not far from that indicated by constancy of angular momentum, or that the circulation approximates that of a true vortex. No satisfactory explanation of the calm center of the hurricane has been made. It is evident, of course, that the true vortex distribution of velocities in the hurricane cannot possibly extend in to the central axis of rotation. This would necessitate an angular velocity inversely proportional to the square of the distance from the central axis, and a linear velocity inversely proportional to the first power of the distance. Long before the central axis was reached velocities would prevail that would require a pressure reduction at the center of the vortex far below anything occurring in the atmosphere. Since these vortices do not extend to the highest atmosphere, and are rather broad relative to their vertical height, this condition would presumably lead to a compensating movement of air down into the vortex, from above, and a partial filling of the vacuum. Apparently this is to a certain extent just what does occur in the calm center, as evidenced by the dissolution of all lower clouds, and occasionally by increased temperature and lowered relative humidity. Probably this adiabatic heating of the central core of the hurricane would be much more marked at some distance above the ground, for at the ground itself little vertical displacement has occurred. The low pressure which does prevail in the center of the hurricane must be attributed simply to the static effect of the potentially warmer air drawn down into the core from above. Hence the higher the cyclone reaches the lower the pressure which it is possible to attain at the center.

In this connection the question of the elevation of maximum intensity of the tropical hurricane becomes important. Presumably, due to frictional effects near the ground, the maximum intensity should be at some distance above the ground. According to V. Bjerknes [38] the suction effect (saugwirkung) of a cyclonic vortex with potentially warmer air above will draw the potentially colder air up in the center from below, the warmer air down from above, toward the level of maximum intensity. The fact that in the eye of the hurricane the lower clouds are dissolved and the humidity falls, is an indication that there is very little lifting up of surface air from below, which indicates that the maximum inten-

sity of the vortex must be low. This is also indicated by the fact that even a very low range of hills is sufficient to destroy a tropical hurricane before it has taken on extratropical characteristics. Indeed in one unusual instance quoted by Hann of a high level station reading close to an intense hurricane (Newera Elya, Ceylon, at 1,890 m., in close proximity to the destructive Backergunge cyclone) only comparatively light winds were reported. On the other hand, cloud observations show that tropical hurricanes extend their influence to at least 8 to 10 kms. Moreover, the fact that they move with the winds at 3 to 4 kms. elevation is a further indication that the center of action of the storm must be well above the ground. In fact, some authorities would have the genesis and development of the hurricane take place at the discontinuity between the trade and antitrade winds as a result of interference between these two wind systems established by strong convection currents to that level. Evidently such questions as these and other problems of the growth and structure of the hurricane cannot be satisfactorily discussed without more data from the upper levels of these disturbances.

For the development of the extratropical cyclone and anticyclone, regional heating and cooling near the ground are of the first importance. This is especially true for the initial stages of the development. As was pointed out in the discussion of the general circulation, the unequal heating or cooling over water and land surfaces is very important in the irregular circulation of middle latitudes, because it establishes longitudinal temperature and consequently pressure gradients. This effect is most marked in the winter, when the poleward temperature gradient is steepest, the general circulation strongest, and cyclones and anticyclones are most numerous. During the cold season, water surfaces are warm relative to land surfaces. Especially is this true where warm ocean currents move poleward to high latitudes. They offer an almost unlimited source of heat and moisture to the atmosphere above, the transfer of heat taking place by conduction, evaporation, and mechanical and convective turbulent mixing. In this way an abnormally warm moist climate is maintained at high latitudes throughout the winter, where the ocean currents are favorable. The two best instances of this occur over the north Pacific, where the Japanese current flows northeastward toward the Aleutian Islands, and over the north Atlantic, from Newfoundland eastward to the south of Iceland and the northern coast of Europe and northward towards Spitzbergen, where the Gulf stream flows. Hence these regions are marked by abnormally low pressure, with some fluctuations, especially throughout the cold season. Such a region of persistent subnormal atmospheric pressure is referred to as a "semipermanent" low, in contrast to the normal "migratory," or moving cyclone. The great semi-

permanent lows are important in the scheme of the general circulation, in that they initiate large-scale northward and southward moving air currents, and are frequently birthplaces of the migratory cyclones.

Corresponding to the semipermanent lows, the result of the heating of the lower atmosphere over a warm water surface, there are the semipermanent highs, or anticyclones, occurring over regions of abnormally great surface cooling for the latitude. Such regions occur in winter over northern Asia and less markedly over the interior of North America. Here the ice- and snow-covered ground loses heat to space by radiation very rapidly, whereas a large part of the scanty insolation is directly reflected to space without absorption. In this way the lower atmosphere is cooled to very low temperatures, the result being a region of prevalently high pressure. At more southerly latitudes cooling takes place over relatively cold ocean surfaces. This is probably the explanation of the Azores High and the South Pacific High. These very extensive anticyclones are portions of the high pressure belt of the horse latitudes, locally intensified by the thermal effect of a relatively cold underlying water surface. Sometimes it is maintained that radiation cooling in the upper levels also plays an important part in the maintenance of these subtropical anticyclones.

Figure 92 represents schematically the above six principal centers of action and the resulting mean winter circulation on the northern hemisphere. It must be remembered, however, that the actual conditions are not stationary as indicated in this scheme. These thermally caused highs and lows initiate air currents as indicated here, but, once started, important dynamic forces come into play. The colder air masses tend to underrun adjacent warmer air masses, hence to move into regions of lower pressure, the warmer air current to overrun the colder air to the north, and the whole system to be displaced eastward with the general eastward current of middle latitudes into which the colder air is usually moving. In other words, the semipermanent high and low initiate air movements which in turn transform the standing cyclone and anticyclone into the migratory types. As soon as this change sets in, these formations can be explained no longer primarily as the result of regional heating and cooling, but rather as a consequence of the dynamic interaction of oppositely directed air currents, and the horizontal transport and displacement of air masses of different temperature and density.

Before passing on to a detailed discussion of the migratory cyclones and anticyclones, it should be pointed out that the source of energy of the initial standing cyclonic and anticyclonic circulation is to be looked for directly in the unequal heating of adjacent regions, *i. e.,* the establishment of strong regional horizontal temperature gradients. A verti-

cal cross-section from the warm to the cold region will indicate a great density of solenoids. The resulting circulation tends to diminish the solenoid density, by so taking place that colder air displaces warmer air at the ground; that is, that the kinetic energy of the system is increased at the expense of the potential energy of the vertical mass distribution. Margules[40] has shown from the computation of quantitative examples for a closed system that this source of energy in the atmosphere is probably frequently quite sufficient to supply the great amount of energy of motion involved in extensive storms or cyclones, as well as to overcome the frictional forces. For this reason the zones of front formation indicated by the heavy lines *AA* and *BB* in Figure 92, the result of the convergent flow of warm and cold air currents initiated by the semipermanent centers of action, are very significant for the growth of migratory cyclones in general, as well as for the appearance of secondary cyclones, or cyclones which develop through the interaction of air currents established by the primary, or initial, disturbance. In terms of the general circulation, these zones represent stretches where the mean tendency is to a sharpening of the polar front discontinuity; the regions of divergent flow *AB* and *BA* represent stretches where the mean tendency is towards a weakening of the polar front and a degeneration of cyclones.

The dynamic and advective factors in the growth and decay of cyclones and anticyclones, low migratory cyclones and anticyclones.—The second cause of excessive or deficient atmospheric pressure at the ground mentioned on page 212 is the mechanical interaction between air currents of different movements and densities; the third is the advective transport of warmer or colder air masses. These two processes will now be considered together. They contribute to the development of the low migratory cyclone and anticyclone.

The discussion in the preceding paragraph indicates how heating or cooling of the lower atmosphere may establish locally regions of low or high pressure. The resulting cyclones and anticyclones are phenomena primarily of the lower atmosphere, having their greatest intensity near the ground where the supply and removal of the heat occurs. This is true especially of the disturbances of higher latitudes, probably less so of the large anticyclones of the subtropics. In their initial stages probably the majority of the cyclones and anticyclones originating at high latitudes are of this "low" variety, though in their later development many of them pass over to the "high" type. At this point, however, the development of the low migratory cyclone and anticyclone, primary and secondary, will be considered.

The gradual formation of an anticyclone by radiation cooling over the interior of a continent and a cyclone by conduction and turbulent heat-

ing over an adjacent warm sea, intensifies locally the poleward temperature gradient, and at the same time establishes a longitudinal pressure gradient which favors a flow of the warm air northward and of the cold air southward. In the case of the high latitude centers of action, this accentuation of temperature contrasts takes place along the polar front, thus furnishing weak spots, or regions of unstable equilibrium, on the front, where the outflow of cold air from the polar regions is bound to occur. Whether this interaction between the cold air to the north and the warm air to the south be considered as a wave motion at a boundary between a denser and a lighter fluid, after the Norwegian school, or simply as an obstructing outflow of the denser fluid into a current of the lighter fluid, after the Austrian school (Exner), makes little difference practically, for any quantitative mathematical treatment of these phenomena appears hopelessly distant at present. Actually Exner's concept seems more applicable to the initial development of the primary or mother cyclone, though in the majority of cases the development of the succeeding secondaries may be traced very successfully to a wave motion initiated on a front greatly sharpened by circulation around the mother disturbance.

According to Exner's barrier theory (Riegel teorie) the development is about as follows: The cold air outbreak towards the south takes place freely between a cyclonic and westward-lying anticyclonic center of action, the normal westward movement which should result from the earth's deflective force being checked by the eastward pressure gradient. Since the warmer air to the south normally constitutes a westerly air current with a slight northward component, the breaking out of the cold air current into this warm air flow effects a certain obstruction or damming up of the current. In the lee of the cold air mass a partial vacuum results, so that the warm air is drawn in from the southwest or south around the cold air mass, the trough of lowest pressure lying between the two currents. This low pressure is caused both by the lee effect of the cold air barrier, and by the normal lightness of the overlying warm air. The following high pressure, or anticyclone, is caused both by the greater density of the cold air in the polar outflow, and by the damming and consequent piling up of the westerly current. This constant force of the westerly current, expressed in the strong eastward pressure gradient, imparts an eastward movement to the cold air barrier at the same time that it is spreading southward and sinking. Thus both the cyclone and anticyclone move first southeastward then eastward from the position of the initial thermally caused low and high, becoming in this way the normal migratory cyclone and anticyclone of the lower atmosphere. Between the warm and cold current there is a discontinuity surface of

the cold front type, while further to the east, where the northward moving warm air current meets the cold air at the polar front, there is a discontinuity of the warm front type. These discontinuities are less clearly marked in the primary cyclone than in the secondaries, therefore the detailed discussion of cyclone structure is taken up under secondary rather than primary disturbances. The kinetic energy of the strong circulation set up during such a development is supplied by the potential energy of the initial vertical mass distribution, or the adjacent warm and cold air masses. As the process continues, the cold air sinks and spreads both southward and eastward, eventually displacing entirely at the ground the warm air current to the east of the cold outflow, *i. e.*, an occlusion has taken place (see next paragraph). From this point the disturbance diminishes in intensity, but the southward transport of cold air in the anticyclone has usually established a very abrupt temperature discontinuity or front at the southern limit of the cold air outbreak. Fronts established in this way are much more abrupt than those which grow up through heating and cooling of adjacent air masses. Unless the condition of stationary equilibrium is rigidly fulfilled at this abrupt front, a secondary disturbance is bound to develop, which under favorable conditions becomes more intense than the weakening primary.

Secondary disturbances. Wave motion, occlusion, and cyclone families.—The sharpest fronts are those established by the southward movement of cold air to some limiting equilibrium position, in the rear of a primary cyclone, as just explained. The abrupt fronts set up in this way sometimes remain almost stationary for periods ranging up to several days. That such a front is extremely favorable for the development of secondary disturbances, which first appear as wave formations on that front is now generally recognized by all who engage in a critical analysis of weather charts. The credit for the discovery of these formations, as well as the general practical analysis of weather condition charts, belongs to the Norwegian meteorologists, particularly J. Bjerknes, H. Solberg, and T. Bergeron.

If a sharp front is to remain almost stationary for a considerable period of time, then the conditions of stationary equilibrium (see page 189) must be very nearly fulfilled. But as soon as any disturbing influence makes itself felt, so that a component of velocity normal to the front is imparted to the air current on either side, then a bending out of the front will appear at some point, and a compensating opposite movement will develop adjacent to it. In other words, a wave is formed on the front. Usually the initial outpushing of the front takes place from the cold into the warm current. The cause of the initial disturbing move-

ment is not always clear. Often it is due to some irregularity in the pressure distribution or air flow quite extraneous to the front itself. Once the disturbance is initiated, the dynamic interaction between the two air currents of different densities and velocities keeps it in existence, and it is propagated along the front. For a wave disturbance once established in this way there are two possibilities; either it continues as a wave formation and a rather weak secondary disturbance which leaves the front intact after its passage, or it occludes, *i. e.,* the cold air flows quite around the center of the disturbance, eventually displacing all the warm air at the ground within the inner portion of the disturbance. This means that this process represents the termination of the stability of the front. In the case of occlusion, the source of the greatly increased kinetic energy of the circulation is the potential energy of adjacent warm and cold air masses. If the disturbance remains a non-occluding wave, the source of the energy necessary for its maintenance is probably to be found in the kinetic energy of the already existing air currents which belong to the general circulation resulting from unequal heating of the earth's surface.

In general, the less the impulse giving rise to a frontal disturbance, that is, the shorter the wave length of the disturbance and the less the velocities perpendicular to the front, the less is the probability of occlusion. It is the waves of small length and small amplitude which remain as waves on the front. The passage of such a wave disturbance is indicated usually only by an inflection in the isobars, and not by a closed set of isobars and complete cyclonic circulation. However, the rainfall attending such a secondary disturbance may be quite considerable, and the relatively small disturbance of the existing pressure field may result in local intensification of the gradient sufficient to cause strong shifting winds. For this reason the prompt detection of such small disturbances is frequently of real importance.

The rapidity of movement of such a small wave disturbance is extremely variable. There are really two components of the velocity of a wave disturbance, that belonging properly to the displacement of the wave along the front through the medium (the air), and that due to the movement of the medium itself (air movement). The first of these, which may be called the dynamic component of the velocity of the wave, is affected by the deflective force due to the earth's rotation in such a way that the wave must move along the front in the direction such that the warmer air is on the right, hence normally towards the east. It is the very long period of these frontal waves which makes the direction of their dynamic propagation dependent on the deflective force, unlike the better known short period fluid waves. This dynamic component of the

wave velocity apparently varies in the normal way for fluid waves, increasing with the wave length and decreasing with the amplitude of the wave, though how closely these variations check up with the theory of wave motion it is impossible to say at present. The movement of the medium itself, or the air currents, is important in determining the net velocity of the wave over the earth's surface. When the air movement and the wave motion proper act in the same direction, very high velocities are found for these wave disturbances. For stable waves of small amplitude velocities well in excess of 100 miles per hour have been noted, which is far in excess of any wind velocities reported within the system. This accounts for the very great speed of movements sometimes noted for minor secondary disturbances. On the other hand, when there exists strong air movement in the direction opposite to the dynamic component of the wave propagation, it sometimes happens that waves of large amplitude remain almost stationary as standing waves on the front, or even suffer a slow retrograde movement. This retrograde movement of a minor secondary disturbance, although not very frequent, may be detected clearly from time to time on the weather map.

In case the wave is unstable and occlusion occurs, the whole history of the disturbance is quite different. If the initial impulse, which started the outflow of cold air, whether due to the instability of the front itself or to a strong external disturbance is sufficiently strong, then the front will not return to its original position. Instead of a wave motion passing along the front there is established a tendency towards a vortical movement about the center of the disturbance, the cold air flowing completely around the center from behind. In that way the original front is advanced far into the region originally occupied by the warm air current. When occlusion is complete, the center of the disturbance is completely surrounded by cold air at the ground, the warm air exists only aloft, and far to the south at the new position of the front.

In this process considerable potential energy of vertical mass distribution is set free. The kinetic energy of the cyclonic circulation, and hence the intensity of the whole disturbance, reaches a maximum at the time of occlusion. Unlike the minor-wave secondary disturbance, a secondary which occludes develops a complete circulation with closed isobars and an intensity which may exceed that of the primary disturbance. The outflow of cold air which occurs upon the occlusion of a disturbance on a front may reach a new equilibrium position, with the re-establishment of the front much further to the south, permitting the development of more secondary disturbances; or it may continue so far that an outpouring of cold air into the trades circulation occurs. When this transport of cold air directly from the polar to the tropical regions has been effected,

the polar front which has been moving southward is temporarily broken, and is subsequently re-formed far to the north on the cessation of the outflow of cold air. The whole sequence of disturbances, from the first thermally formed cyclone and anticyclone in the far north, the occlusion of the primary cyclone and consequent accentuation of the polar front with the subsequent secondary disturbances, occluding or non-occluding, down through the final occluding disturbance which terminates the series with the outflow of cold air into the trades—such a complete series of cyclones J. Bjerknes has called a cyclone family. The entire series occurs on one continuous section of the polar front separating a vast warm current of tropical origin on the one side from the cold current of polar origin on the other side. Usually these two currents move eastward at the same time that the cold outflow, and consequently the polar front with its successive disturbances, is moving southward. The polar front itself lies along the trough of minimum pressure between the two air currents. This trough is the line followed by the successive waves of high and low pressure as the front advances and recedes with the passage of successive disturbances. The advance of the cold current is marked by rising pressure and the establishment of anticyclonic conditions. The only published thorough and analytical treatment, and that an exceedingly instructive one, of a particular instance of the formation of a sharp front, the development of successive wave and vortex disturbances, and the termination of the cyclone family, is the well-known work of Bergeron and Swoboda.[41]

Before leaving the subject of wave formations on a front, a brief discussion of the distribution of the meteorological elements about such a disturbance is required. Figure 93, after J. Bjerknes and H. Solberg,[42] represents horizontal and vertical cross-sections through an ideal non-occluded disturbance (open warm sector) and through an ideal occluding wave disturbance. Figure 93(a) shows the distribution of elements and the position at the ground of the fronts in the case of the open warm sector, while 93*b* shows a vertical cross section through the same disturbance along the line *AB*. In Figure 93(b) the typical warm and cold front discontinuities with their characteristic condensation forms and precipitation (see page 193) will be recognized. Thus the warm sector of a young and active cyclone is preceded by a lowering cloud deck and a broad precipitation zone, with a flow of cold air almost parallel to the advancing warm front and a rapidly falling barometer. The passage of the warm front is marked by a cessation of steady rain, a shift of wind by perhaps 90°, a rise in temperature, and sometimes by partial or total clearing. Clouds, if present, are usually low and stratiform, the barometer falling slowly. The passage of the cold front is marked by an abrupt wind

shift, frequently by a squall or heavy but rather brief shower, colder air, and rapidly rising pressure. Usually clearing follows rapidly, only cumulus clouds remaining, unless the exposure is maritime, when scattered convective showers may continue for some time. The winds about

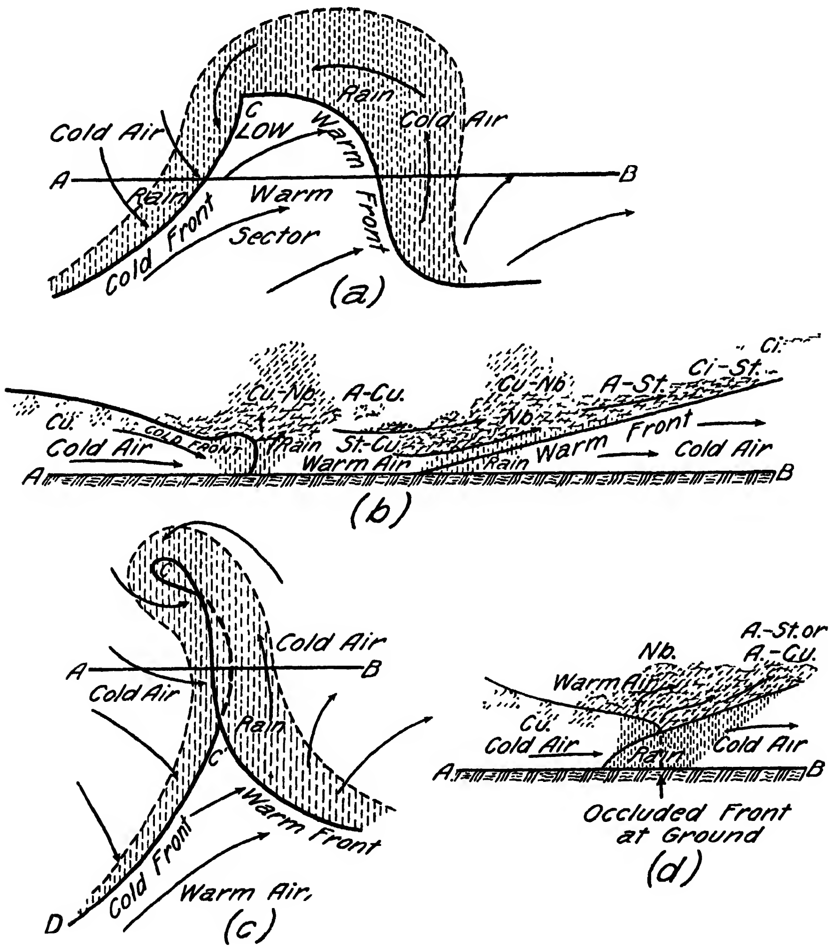

FIG. 93.—Horizontal and vertical cross-sections of open and occluded warm sectors.

the disturbance are indicated by the arrows, the main zones by shading. The center of the disturbance is at C.

Figure 93(c) shows the horizontal section at the ground of a typical wave disturbance which has just occluded, and 93(d) the vertical section through the same along the line AB. It will be noted that in the central part of the disturbance the cold air from behind has overtaken that in

front, or that there is only one front line, wind shift line, and rain zone. With the passage of this front the wind shifts by more than 90 degrees. After the occlusion, with the continued lifting of the warm air, the intensity and breadth of the rain zone and the cloud area steadily decrease. It frequently happens that the central portion of the disturbance, at C, is the last to occlude, so that there may remain a remnant of warm air at the ground there after it has disappeared from the ground further from the center. The occluded front is preceded by falling pressure, and followed by rapidly rising pressure, but normally after occlusion the intensity of the disturbance, at least at the ground, steadily diminishes. Further secondary disturbances are likely to develop either at C', where there remains the remnant of a warm sector, or as a wave disturbance along the cold front out towards D. With the continued lifting of the warm air at the old center C, the level of maximum intensity of the disturbance is progressively raised. Therefore new-formed disturbances as a rule have open warm sectors and maximum intensity near the ground. Correspondingly, the accompanying anticyclone is a low anticyclone, due to the advance of a cold air mass in the lower atmosphere. This does not mean, however, that the cold air current or polar outflow may not be a rather deep current, but only that the maximum intensity of the anticyclone is near the ground. The deeper the current, the more intense will the anticyclone be. Hence in regions of frontogenesis, as from central North America eastward over Newfoundland, where the migratory cyclones develop so markedly, these cyclones are likely to be predominantly of the low type, or with maximum intensity near the ground. On the other hand, in regions of frontolysis, as from the central north Atlantic eastward into Europe and Asia, where the cyclones are mostly filling (warm air being displaced by cold at low levels) disturbances with a long past history, they are likely to be of the high type, or with maximum intensity in the upper levels of the troposphere. However, to explain the development of high cyclones and anticyclones it is necessary to consider the possibility of advection of warm and cold air masses at high levels.

Dines' and Schedler's correlation coefficients—Possible explanations—Transition from low to high cyclones and anticyclones.—Up to the present, the discussion of cyclones and anticyclones has been directed primarily toward the disturbances whose greatest intensity occurs in the lower troposphere, the low cyclones and anticyclones, whose genesis and maintenance are to be looked for in the sharp horizontal temperature contrasts characteristic of certain regions near the ground, and in the dynamic interaction as wave or vortex motion between large-scale air currents initiated by these temperature contrasts. Although data ob-

tained by soundings of upper air conditions are exceedingly scanty compared to the extensive material at hand for the study of surface conditions, there is enough upper air material in certain situations to indicate that changes in the temperature and pressure distribution at the base of the stratosphere occur, which are strikingly similar to the cyclones and anticyclones of the lower troposphere. The existence over western Europe of disturbances having their maximum intensity in the substratosphere has been shown conclusively by analytical methods notably by Rossby and Haurwitz, and less satisfactorily, due to less satisfactory data, for North America. Statistically not only their existence, but their close connection with surface conditions over western Europe has been proved by the interesting correlation coefficients of Dines and Schedler.

By using data collected over western Europe during one of the international upper air sounding periods, Rossby [43] was able to show not only the existence of closed isobars at the 12 km. level, but also an asymmetrical distribution of temperature about the center of a disturbance corresponding exactly to the warm sector of a non-occluded cyclone of the lower troposphere, the warm sector being displaced by much colder air by the following day. In another instance a region of low pressure filled and warmed uniformly throughout, the entire change evidently being dynamically caused, by an inflow of air, compression, and adiabatic heating. Such analytical studies of individual instances of changes in the substratosphere indicate that both advective and dynamic factors are at work there as in the troposphere.

Dines [44] took a great number of individual upper air soundings over western Europe and worked out the correlation coefficients between the following quantities:

1) Air pressure at the ground
2) Temperature at the base of the stratosphere
3) Elevation of the base of the stratosphere
4) Pressure at the 9 km. level
5) Mean temperature of the lowest 9 km. air column.

Schedler [45] worked out the same coefficients for series ascents, *i. e.*, for soundings made at the same station on successive days. The following tabulation, taken from Exner,[46] gives the correlation coefficients between the above five quantities, as determined by Dines and Schedler. In each case the two-digit subscript gives the two factors being correlated:

	Dines	Schedler		Dines	Schedler
$r_{1,2}$	−0.52	−0.38	$r_{2,3}$	−0.68	−0.47
$r_{1,3}$	0.68	0.29	$r_{2,4}$	−0.47	−0.33
$r_{1,4}$	0.68	0.45	$r_{3,4}$	0.84	0.73
$r_{1,5}$	0.47	....	$r_{4,5}$	0.95	0.87

Especially striking is the high positive correlation between factors 3), 4) and factors 4), 5). They show that high pressure at 9 kms. goes with a high stratosphere, which means also a cold stratosphere, but with a warm troposphere. Coefficients $r_{1,3}$ and $r_{1,4}$ show further that there is a marked tendency for high and low pressure at the ground to be accompanied by high and low pressure at 9 kms., and hence with a high cold and a low warm stratosphere, respectively. This is in accord with the previously mentioned fact that the stratosphere tends to be lower and warmer over cyclones, colder and higher over anticyclones, whereas most of the troposphere above the surface is colder in cyclones, warmer in anticyclones. In other words, statistical evidence shows that in the majority of cases the pressure variations observed as cyclones and anticyclones at the earth's surface have their maximum intensity at about 9 kms., and that it is the stratosphere which contributes the deficit or excess of air mass which is indicated by these phenomena. This is strikingly confirmed by Schedler's analysis of the contribution of the different kilometer levels to the net 24-hour surface pressure changes on days of rising and of falling pressure. He finds that in the mean the first 3 kms. contribute to a small extent directly to the observed surface pressure change, and that the next 5 of 6 kms. act weakly against it, so that at 9 kms. the changes average numerically as large as the surface changes, and relatively much larger. About 40 per cent of the surface change is contributed by the layer from 9 to 14 kms., and the remaining 60 per cent by the stratosphere above 14 kms. This means that over Europe relatively few of the large pressure changes are due to air mass displacements at the ground, but that the cyclones and anticyclones are prevailingly of the high type. All this is in accordance with the fact that the continent of Europe lies in a belt of divergent flow in the general circulation and consequent dissipation of fronts and cyclonic disturbances. The cyclones that move in over Europe are usually long occluded remnants of formerly much more intense surface disturbances, in which the center of action has been continuously displaced to higher levels. Similarly, the anticyclones or pressure rises are either due to successive eastward displacements of the great semi-permanent Azores high, or are of the large slowly growing and nearly stationary type which frequently develop gradually from anticyclones of the surface cold air type, probably through the cooperation of the stratosphere. In Europe there is no rapid succession of cold and warm waves of great temperature contrast, and the accompanying active young developing cyclones and anticyclones of the low type, so characteristic of the region from the central United States northeastward over the north Atlantic, a region of convergent flow and of front formation in the

general circulation. Probably an analysis such as Schedler's would show a much greater predominance of low-level action in the United States than in Europe.

Explanation of changes which take place in the substratosphere.—The changes which take place in the substratosphere have been explained in different ways. Some of the more important will be considered briefly at this point.

1) *Advection.*—That the stratosphere becomes warmer and lower from the subtropics towards the poles is a well-known fact. There exists, then, in the substratosphere an equatorward temperature gradient which is only slightly less marked than the poleward gradient in the troposphere. Although there probably exists no such abrupt transition zone as that of the polar front in the troposphere, the term ***equatorial front*** is applied to the zone of most rapid transition, and it seems very probable in view of the analytical work of Rossby and Hourwitz mentioned above, that wave disturbances and dynamically interacting oppositely directed currents may exist on this front as in the troposphere. Since very high wind velocities are known to exist at these levels, great latitudinal displacements may take place in a very short time, hence extensive cold and warm air transport may markedly change the local stratospheric contribution to the surface pressure.

2) *Radiation.*—Lowering of the stratosphere temperature and consequent raising of the elevation of its base will follow from the interception of part of the long-wave radiation passing out to space from the earth's surface before it reaches the stratosphere. Therefore an increase in moisture in the troposphere will result in a colder higher stratosphere, a decrease of moisture in a lower and warmer one. This accounts for the lowering of the stratosphere from equator to pole, and, according to R. Mügge,[14a] in a large measure for both the semi-permanent and the migratory high anticyclones. It is possible that this may account in part for slow changes in the stratosphere, but probably not for the more rapid changes, and certainly not for cyclones with asymmetric temperature distribution.

3) *Saugwirkung* (*suction*).—An entirely different line of thought is that developed by V. Bjerknes, that the lifting and sinking of the discontinuity surface at the base of the stratosphere, with consequent adiabatic cooling and heating of the air above, is simply the dynamic consequence of the existence of unequal rotation in the anticyclonic and cyclonic sense above and below the stratosphere inversion (see page 191). This theory, however, makes both the temperature and pressure distribution the consequence of the state of motion of the atmosphere, hence in no way contributory to that motion, nor does it permit of an explanation

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CHAPTER VI

PHYSICAL BASIS OF WEATHER FORECASTING

Richard Hanson Weightman

INTRODUCTION

All weather phenomena are the results of processes due to the operation of physical laws. Meteorology, in spite of the wonderful progress that has been accomplished in the last decade or so, will probably never become in its entirety what is commonly known as an exact science. Without doubt, however, certain phases of meteorology at least will attain that end, but the manifold complexity of the problems demands and will continue to call for extensive further development of both theoretical physics and observational meteorology. Discussions of meteorological facts as coordinated with physical laws are of the utmost importance to the practical problem of forecasting, since they help to a better understanding of the phenomena of the atmosphere, and the processes that effect weather changes. Furthermore, in addition to having scientific importance and interest, they bring us nearer the ultimate goal of making forecasting a well rounded-out science instead of an art.

The art of forecasting developed through empirical rules has led to excellent results in the hands of the experienced, but on such basis its capacity to further develop is limited. To carry it further or to make out the true inwardness of its application in special cases, we must depend upon a knowledge of the dynamics and physics of the atmosphere.

With complete observations available from an extensive portion of the free air, the problem would be to apply the equations of mathematical physics to the actually existing atmospheric conditions and to compute the conditions to follow. Such calculations would, of course, require a great amount of labor and time, but progress can be and has been made by simplifying assumptions in calculating certain phases of the synoptic charts, whereby good qualitative results are obtained but not such good results as to quantity, for which we still have to depend mainly on experience.

Bjerknes [1] developed the first plan of predicting weather mathematically. It seemed obvious to him that the usual mathematical methods could not be adapted to a problem of this sort, and that there could be no thought of an analytical presentation of the observational results with a subsequent analytical integration of the equations. As the observations

are presented by means of charts, all mathematical computations must be recast into graphical operations by means of maps. In this way, he said, we have developed for ourselves the rudiments of a graphical mathematics by means of which we derive one map from the other, just as one derives one equation from another by calculation.

Later, Richardson [2] discussed the motions and phenomena of the atmosphere under the combined influences of all the principal factors, including radiation and absorption, evaporation and condensation, eddy motions and turbulence in the lower atmosphere, gathered them into one set of systematic mathematical equations, and attempted to utilize them in computing future conditions. He employed seven dependent variables and the four independent variables; time, height above mean sea level, longitude and latitude. Atmospheric phenomena were then completely described by seven fundamental differential equations, in the solution of which the method of finite differences was employed. The method was applied quite satisfactorily to a very complete set of observations over middle Europe for May 20, 1910, for which Bjerknes had published detailed charts.

The usual objection to such mathematical treatment is "How can this be of use? The calculations require a long time, measured by weeks or months." To this question Bjerknes' reply, which has become almost classical, was "If only the calculation shall agree with the facts, the scientific victory will be won."

While this reply may cleverly answer the usual objection, it is a fact that as yet forecasting in its entirety, or even for the greater part, presupposing the availability of sufficient data, cannot be accomplished expeditiously by mathematical methods.

Modern forecasting still remains empirical in large measure. However, with the increasing knowledge of the structure of highs and lows, the processes concerning their genesis, continuance and dissipation, and the laws of precipitation causation, forecasting is slowly but steadily developing into a quasi-exact science at least, instead of remaining an art. The impetus to progress in this direction has come from the increasing number of observations in the free air by means of pilot and sounding balloons, kites and aeroplanes.

1. HIGHS AND LOWS

At the earth's surface, owing to friction, winds cross the isobars at an angle averaging about 30° over land and about 20° over water, while the velocities are fairly proportional to the pressure gradient. In the free air, at an elevation generally taken as 500 meters, the gradient-wind relation is that the direction of the wind is tangent to the isobars and the

velocity proportional to the pressure gradient. Therefore, if the pressure distribution is known, the wind circulation is also known, at least for the free air, so that the general problem of forecasting may be said to be reduced largely to a pressure forecast problem; that is, if the pressure distribution twenty-four hours ahead can be forecast, the distribution of winds will also be known.

The effect of the rotation of the earth as shown by Coriolis and elaborated by Ferrell is the same as if every portion of the moving air were subject to a force across its path from left to right in the northern hemisphere equal to $2\omega\rho V \sin\phi$, where ω is the angular velocity of the earth, ϕ the latitude of the place, V the velocity of the air and ρ its density. Therefore, for motion in a great circle the air must be pushed in the opposite direction by a force across its path to balance the deviating effect of the earth's rotation. The pressure distribution will supply the necessary force if its gradient γ is of the right value; hence the equation $\gamma = 2\omega\rho V \sin\phi$ gives us the condition under which the air will keep exactly straight. But generally in meteorology we find in combination with the directive action of pressure due to the rotation of the earth, the centrifugal action around a center of low or high pressure. In such cases we have $\gamma/\rho = 2\omega V \sin\phi \pm \frac{V^2}{R} \cot a$, the last term being positive with LOWS and negative with HIGHS. Shaw gives a table in his Forecasting Weather showing distance apart in nautical miles of consecutive 1/10 inch isobars in latitude 53° corresponding with stated gradient velocities, pressures and temperatures; and elaborate tables are given by Humphreys in his Physics of the Air for both cyclones and anticyclones. Such values apply to the free air and not to surface conditions, the latter being greatly modified owing to friction and topography, varying greatly in numerous cases for stations quite close together.

In drawing on weather maps isobars or lines of equal pressure reduced to sea-level or some other common plane, there are found regions of high pressure and regions of low pressure, the pressure diminishing from the center of high to the center of low pressure. Cyclones, or "lows," may be described as wind systems in which the instantaneous direction of the surface air is inward, counter-clockwise (in the northern hemisphere), at an appreciable angle across the isobars toward the center of lowest pressure. Such movement of the air is due to the combined effects of the deflective force of the earth's rotation and the pressure field. For determining the general direction of the low center with reference to the observer, a rule enunciated by Buys Ballot of Utrecht is useful. The rule for the Northern Hemisphere is generally stated "Stand with your back to the wind, and the barometer will be lower on your left hand."

The reverse is true in the southern hemisphere. This rule does not, of course, permit the definite location of the low pressure center for three reasons; First, the LOW, except in rare cases, does not have circular isobars; second, pressures and surface winds are not symmetrically distributed around the low center; and third, the surface winds are not tangent to the isobars, as they are in the free air. The amount of incurvature at the surface varies, but averages about 30° over land and about 20° over water, as illustrated in Figure 94.

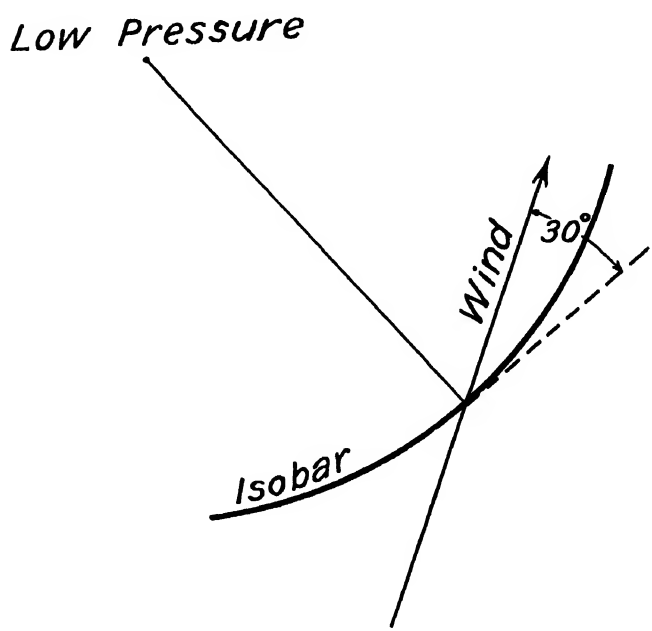

Fig. 94.—Wind direction in relation to isobars.

On the average, there is more deviation of the wind from the tangent to the isobar in the HIGH than in the LOW. See Figure 95.

LOWS as well as HIGHS are relative phenomena. There may be a LOW with central pressure of 30.20 inches or a HIGH with the same central pressure, all depending on the relative pressures surrounding them. HIGHS in the northwestern United States occasionally show pressures at the center exceeding 31.00 inches and in LOWS the pressure is depressed occasionally to below 29.00 inches. In the storm of March 7, 1929, a LOW was central north of the Canadian Maritime Provinces, near Belle Isle, with an unusually low pressure of 28.14 inches at that station.

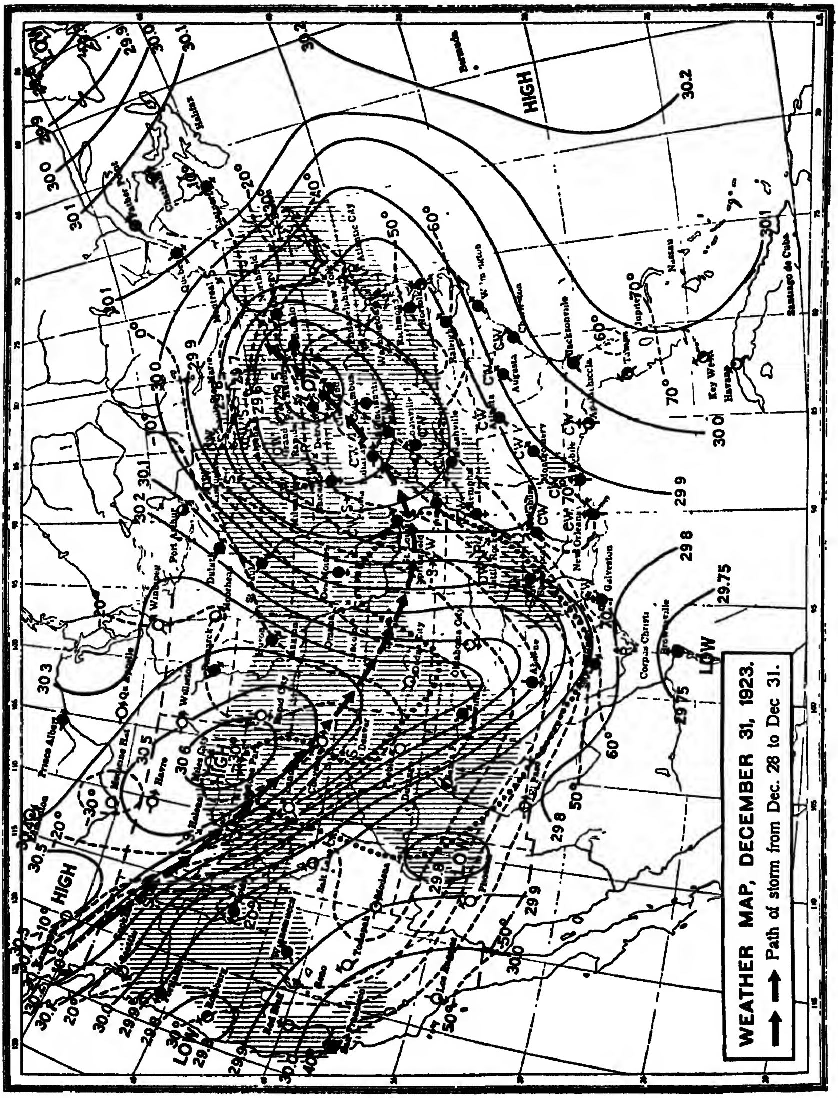

FIG. 95.—Weather map. Dec. 31, 1923 (8 a. m.).

A. Classes of Lows.—*Semi-permanent and migratory.*—LOWS may be divided into two general classes: 1) the semi-permanent or seasonal, and 2) the migratory. The principal semi-permanent LOWS in the Northern Hemisphere are the summer LOWS over Siberia and the interior of the North American continent and in the winter time over the Aleutian Islands and southern Kamchatka. The only one of this kind that is in evidence both summer and winter is over and to the southwest of Iceland, in which the lowest average monthly pressure is about 29.60 inches, during the month of February. Less important LOWS accompanied by wind systems are in evidence over interior Alaska and Spain during the summer and over the Caspian Sea in winter. None of the LOWS referred to above are permanent for they give place at times to anticyclonic conditions. Prof. A. J. Henry has termed them " Statistical LOWS " which expression, suggesting as it does the thought of averages, which may be made up of high readings mixed with predominatingly low pressure readings, well expresses their character. It should be remarked, however, that the conditions that are responsible for their occurrence persist all the time.

The migratory LOWS are ones that move or migrate generally from west to east in the Northern Hemisphere at the rate of 500 to 700 miles per day. In the migratory class may also be included the tropical cyclones which, during their journey in the tropics move at first from east to west but after they recurve into temperate latitudes are carried along in the west to east circulation.

Tropical and extra-tropical.—LOWS may also be distinguished as tropical and extra-tropical depending on their region of origin. Tropical LOWS are either very flat, shallow disturbances giving rise to rains but without winds of consequence or they may be the violent storms denominated " Hurricanes " in the West Indies, " Typhoons " in the East Indies, " Baguios " in the Philippines, and " Cyclonic Storms " in the Bay of Bengal and Indian Ocean.

Although tropical and extra-tropical storms have some common characteristics, there are important differences. Humphreys in his Physics of the Air has set out a number of them, the most important of which are: 1) The isobars of the tropical cyclone generally are more symmetrical and more nearly circular than those of the extra-tropical. 2) The temperature distribution around the vortex of the tropical cyclone is practically the same in every direction while about the extra-tropical it is very different. 3) The rainfall in tropical cyclones is heavier and more symmetrically distributed than in the extra-tropical. 4) Tropical cyclones are most frequent during the summer of the hemisphere in which they occur, whereas the extra-tropical are strongest and most nu-

merous during the winter. 5) Tropical cyclones often move to higher latitudes when they assume, more or less completely, the characteristics of the extra-tropical; the extra-tropical, on the other hand, never invade the region of the tropical nor assume its distinctive characteristics. 6) The tropical cyclone has no anticyclonic companion, while the extra-tropical has one in practically every case.

According to form of isobars.—The class with which we are most concerned is the migratory, extra-tropical LOW, for it is these LOWS that bring to the region over which they pass the successive changes in weather characteristic of temperate latitudes. Such LOWS have a variety of forms: the round LOW, the trough type (which includes the "V"-shaped isobar type, the inverted "V" type and the elliptical form), and the col or saddle. These types may be found in Abercromby's classification.[3]

The round LOW.—This type, in which the isobars are roughly circular, is manifested in hurricanes, but even in these the isobars are not truly circular except, possibly, very close to the center. In extra-tropical LOWS, isobars have a tendency to become roughly circular in the U. S. after the LOW has moved northward with a considerable increase in intensity.

The trough type.—The trough type includes LOWS of the "V" type, the inverted "V" type and the elliptical type. In the "V" type the central isobars, of which there may be several, are more or less circular, but the isobars next outward from them have the form of a "V." In such cases HIGHS are generally situated to the west and to the southeast of the LOW center. In the inverted "V" type, the isobars may be similarly described, except that the pressure system is inverted in a north and south sense, the HIGHS in such case being to the northeast and northwest of the low center. The elliptical type is one in which the isobars are oval, the major axis being to the minor axis in the ratio of 2 or more to 1. There may be a center in one end of the oval, or there may be a center in each end and occasionally a third center in the middle. The outer isobars on the east and west sides frequently trend nearly north and south and may be approximately straight. A system with several centers does not retain its form for any length of time, because one of the centers deepens and the whole system changes shape.

The "col" or "saddle."—This type is composed of a system of isobars involving a LOW to the south saddled by HIGHS, one to the northeast and the other to the northwest of the low center. In some cases a LOW may be to the north of the first low center, but in either case pressure is relatively high between the two high centers. This type is sometimes confused with the "Wedge" in which a neck of high pressure from a

HIGH to the southward separates two LOWS, one to the northeast of the HIGH and the other to the northwest. In the latter case the HIGH is improperly likened to a saddle on which the LOWS rest, whereas in the former, the HIGHS are considered the saddle, resting on the horse's back in the shape of a LOW.

B. Classes of Highs.—Anticyclones or "HIGHS" are wind systems always of extratropical regions in which the instantaneous direction of the surface air is outward clockwise (in the northern hemisphere) from the center of highest pressure.

The seasonal HIGHS of the northern hemisphere are central in winter over Siberia and the interior of the northern United States and Canada, while HIGHS in the region northeast of the Azores and northeast of the Hawaiian Islands are in evidence over the oceans throughout the year. These HIGHS are all semi-permanent in character and at times give place to cyclonic conditions.

Migratory HIGHS have a variety of forms as judged from the shape of isobars. Some are round, others elliptical, while some have the form of a "V" oriented in a variety of ways.

So far the air circulation attending LOWS and HIGHS has been described only as regards surface conditions. Later, in Section 2-D, under the "Norwegian Method," their vertical structure will be discussed.

HIGHS and LOWS in the United States have been classified by regions of origin [4] or first appearance, because each type has important characteristics which persist during the passage of the cyclones and anticyclones across the United States. For example: LOWS coming from Alberta have quite different effects from those coming from the Pacific or Texas. Also, HIGHS of continental origin are more intense, though shallower, and carry colder weather farther south than those of ocean origin coming from the Pacific.

C. Origin of Lows.—Theories of the origin of LOWS have undergone many changes. The *convectional theory,* with local heating as the basis, gave place to the *counter-current theory* which involved the mechanical development of a vortex between two oppositely directed currents, with a northward moving current to the east of a southward moving current, or an eastward moving current to the south of a westward moving current. It was Margules who pointed out that the most important consideration was the difference in temperature and consequently the density difference between two air masses flowing side by side in opposite directions or in the same direction with different velocities.

Margules has also shown that, should anything tend to destroy the balance, when two air currents situated as indicated in the first example are in dynamic equilibrium, then the colder will tend to underrun the

warmer, and a cyclone will be produced. Such a condition, undoubtedly, constitutes the basis of origin of most of our LOWS.

Margules further pointed out that whatever theory is evolved regarding the formation, maintenance and decay of cyclones, there are four vital features that have to be explained: (a) the propagation or travel of the disturbance; (b) the removal of the rising air; (c) the supply of energy, and (d) the circulation or air-current structure. Perhaps the most important is the supply of energy, which has been discussed by Margules, Exner, Ryd and others.

Another theory developed by the British deals with cyclonic disturbances as vortices. They may not often manifest themselves in nature as such, but if we consider the modifications due to the effect of friction and movement of translation, they may be recognized, certainly at times at least, in our cyclones, especially those that deepen rapidly. For the initiation of disturbances it is necessary, according to Shaw,[5] to presuppose a condition of vorticity, that is, involving incipient and probably irregular vortices, over an area. When such a condition exists, it has been shown by Rayleigh that, if eviction of air takes place (in some upper level), a principal vortex will develop in the area, absorbing the incipient vortices. Investigations along similar lines have also been pursued by Kobayasi and Fujiwhara.

The vortex idea is quite graphically brought out if the observed pressure field is separated into two component fields, the first, which is called the stationary field, consisting of parallel isobars, and the second, the impressed field, consisting of circular isobars.

This separation into two fields is also used by Ryd in his Traveling Cyclones[6] and by Fujiwhara[7] and Kobayasi.[8]

Bjerknes[9] has offered the "polar front" theory which attributes the source of energy to the occasional outbursts of cold (polar) air from northern regions which result in the development of LOWS along the boundary between polar and equatorial air. The cold and warm currents may exist side by side in dynamic equilibrium until something causes a disruption of the balance, when an invasion of colder air into the more southerly regions takes place causing the formation of a LOW. The cause of this disruption Bjerknes attributes to waves of Helmholtzian nature developed along the boundary between the warm and the cold currents.

In the "drop theory" of the Austrian School[10] such disruption of balance is caused by mountain ranges or other physical obstacles. The Norwegian ideas involve an unbroken surface of discontinuity around the hemispheres between the polar and equatorial air masses, whereas the Austrian ideas do not require this, which latter are more in harmony with observations.

2. FORECAST METHODS

A. Based on Movement of HIGHS and LOWS.—The earlier forecasters based their predictions on the movements of HIGHS and LOWS as indicated by surface isobars and the conditions that attended them. Such knowledge was gained through a careful study of many weather maps. Unfortunately a great deal of such knowledge has not been published and many excellent maxims, no doubt, have been lost. However, a number of rules gathered together by Bowie,[11] are preserved in printed form and a wealth of useful knowledge is to be found in Weather Forecasting in the United States.[12]

First, the forecaster must be able to predict the general pressure distribution and estimate closely the subsequent positions of the HIGHS and LOWS as well as the troughs and wind shift lines attending the latter—otherwise it is impossible to predict the winds and changes of temperature, and in an equal degree the precipitation. In 1906, Bowie announced a method,[13] for determining the movement of a LOW by taking into consideration the pressure distribution around the low center. With practice in the use of this method satisfactory results can be obtained in most cases. Another indication that has been found useful is that the LOW in most cases will follow the isotherms during the succeeding 12 hours at least, and for 24 hours when there is a marked temperature gradient in the region of and to the north or northwest of the low center, especially with isotherms running nearly east and west. It is well also to consider the previous 12- and 24-hour movement of the LOW, both as to direction and rate of travel. Further, it is important to note whether the LOW is deepening or filling up. The pressure map showing 12-hour changes, together with the tendencies or pressure changes in the two hours preceding the observation should also be carefully considered, attention being given as to whether the maximum 12-hour fall is greater or less than at the preceding observation. While some have been led to believe that the low center will move toward the region of greatest 12-hour pressure fall, this is seldom true, nor should it be expected for reasons which have been so clearly pointed out by Bowie.[14] In considering the direction and rate of movement of a LOW, it is also of importance to note the movement of the preceding HIGH, whether it is increasing or decreasing, or changing direction, together with the pressure changes attending it. A fast movement of the preceding HIGH generally indicates a fast movement of the LOW following, and vice versa.

At times when a large HIGH is stationary over the Canadian Maritime Provinces the LOW over the Lake Region will stagnate, *i. e.,* make little if any eastward progress. Professor Garriott called attention to

the fact that stagnation over the Canadian northeast was preceded about six days by stagnation over the northern British Isles and Iceland, and, further, that about six days after movement was resumed over Iceland, normal movement followed over the Newfoundland region. The normal winter rate of movement of LOWS over the northern route is about 700 miles per day, but frequently during periods of a week or ten days at a time, the rate of movement will be 1,000 to 1,500 miles a day. Once such rates of travel become established, allowance has to be made for them. When the rate of eastward drift changes slowly it is not a difficult matter to estimate such movement, but when the rate of progress suddenly increases or decreases, the ability of the forecaster is taxed.

The great majority of LOWS in the United States are of the trough type of which there are a number of variations, the principal three of which are 1) trough with principal center in the northern end, commonly known as the "V" type; 2) inverted "V" type with principal center in the southern end; and 3) the elliptical type which may be of the broad or the very narrow form and in which there may be several centers. If a LOW develops a trough, it almost always decreases its rate of movement. If a LOW'S trough is cut off, or occluded, the rate of movement of the LOW will be accelerated. In studying troughs it is very important to determine whether the axis of the trough is changing direction. If the northern end is moving faster than the southern, such condition is usually followed by unsettled weather, and by the development of another LOW somewhat later in the southwest. If the southern end is moving faster than the northern, then there is almost an even chance that a secondary disturbance will develop quickly in the southern part of the trough, while the center in the northern end will decrease in intensity, followed by settled weather, at least in southern districts. A study of a number of cases of the latter type indicates that the first definite indication of the development of a secondary is a localized 12-hour pressure fall in the southern part of the trough associated with a definite local cyclonic wind circulation. Without this there is little chance of an immediate development.

After estimating the direction and rate of movement of the LOW and its attendant trough, it is quite important to determine whether it will increase or decrease in intensity, for which few satisfactory precepts can be given based on surface conditions alone.

The occurrence of precipitation has to be classified according to numerous climatic regions for purposes of study, and it would be beyond the scope of this paper to attempt here to give even a majority of such rules, especially in view of the fact that a more generalized method will be discussed later in the section under "The Norwegian Method."

B. Weather Catalogues.—Useful weather catalogues, for the most part unfortunately not published, have been compiled by different forecasters as aids in their work. Many of them were directed to the occurrence of conditions at local stations, such as pressure types under which frost, cold waves, heavy snows, and precipitation will occur. Considerable information is to be found in Weather Forecasting in the United States. A discussion of a general nature was published by E. Gold,[15] covering the general pressure types for the British Isles, along lines suggested by Abercromby.

It is realized, of course, that the most probable or average movement of a certain type of storm cannot represent all individual cases and will only aid in a general way. The average can only serve as a general guide, and the behavior of the individual case becomes the forecaster's problem. Although such aids have been extremely useful, experience has taught us that it is only by a painstaking study of the weather charts that satisfactory forecasts can be made.

With further regard to local weather catalogues, even if the HIGHS and LOWS have the same location and are approximately of the same intensity as in past cases, there is no assurance that the results will be the same as in the case under immediate consideration, because the LOWS and also the HIGHS may have totally different origins and directions of movement. If such is the case, we certainly should not expect the same results to follow. Among those who have devised weather catalogues or indexes but did not publish them are Brandenburg, J. Warren Smith, Brist, Henry and others. Of a somewhat different type are those indicated by Rolf[16] and Blair.[17] In such studies graphs are worked out showing the probability of occurrence of precipitation, based on the different elements, such as pressure, wind direction, humidity, etc.; also, on changes in such elements and combinations of both the elements and changes.

C. French Methods.—For years the different meteorological offices have given attention to the changes in pressures as well as to the pressures themselves. Pressure changes have been studied and charted for over fifty years in several countries, including the United States. The first writers on the subject were Eckholm, Brounow and Svrensky. In former years some of the meteorological services used the 24-hour and 12-hour changes. For some years these have been supplemented in Europe by the "tendencies" or changes within the three hours preceding the observations (two hours in the United States). In Europe where four observations a day are telegraphed, it is possible to construct charts approximately six hours apart, and with the aid of the tendencies (or change within the three hours preceding these observa-

tions) to construct charts approximately three hours apart, a total of eight in 24 hours, if desired.

Pressure change.—A method developed during the World War by the French closely associates the pressure change areas and the cloud systems, but has not been followed to any appreciable extent as such, except in that country. However, it is sufficiently interesting to present some of its features here.

In commenting on the subject, the French [18] have pointed out quite logically that the 12-hour pressure changes prepared at 3-hour intervals show the change in and development of the barometric situation more quickly and accurately than the 12-hour changes prepared only twice a day based on the 7 a. m. and 6 p. m. observations.

The following ideas from the French system of pressure changes apply more particularly to weather of the westerly type, and only when quite well-defined LOWS and a well-developed HIGH are involved.

It is stated that all 12-hour change areas move at appreciably the same rate, 700 km. in winter and 500 km. in summer in 12 hours, depending only on the season, and that the amount of maximum change accompanying a particular depression does not vary greatly during its passage over western Europe, even though the depression may increase or decrease. It is further stated that the 3-hour changes are approximately equal to one-third of the 12-hour, and the 12-hour change equal to two-thirds of the 24-hour change.

To estimate the previous position of the 12-hour maximum pressure change and to compute the direction of movement of the pressure change area, when the change is making its appearance on the western border of the map, the following method is used: Take for example the 18-hour map, as the map from which to work, and make a 12-hour (11-hour) pressure change chart by comparing the 7-hour with the 18-hour pressure. By applying the appropriate tendencies to the 7-hour and 18-hour pressures, the 4-hour and 15-hour pressures may be obtained, from which may be constructed also a 12-hour (11-hour) change map 4 hours to 15 hours. Take a single station and note the 12-hour change on 15-hour and 18-hour maps, which are three hours apart. For example, on the 15-hour map the rise is 7.7 mm., and on the 18-hour map 4.7 mm.

Assuming that the shape and spacing of the allobars do not vary, an arc is drawn, using the station as a center, and taking the distance moved in three hours as 3×60 km. $= 180$ km. as a radius. The point of intersection of the arc and the 4.7 mm. line of 12-hour change on the 15-hour map will fix a point over the ocean where the pressure change was 7.7 mm., three hours previously. The process is repeated for all stations, thereby projecting backward the 12-hour pressure change over the ocean, and lo-

cating as well the center of maximum fall or rise on the 15-hour map. Drawing a line between the center of maximum fall or rise on the 15-hour and that on the 18-hour map, gives the direction of movement of the change area.

These rules, and others that follow from them, are applied in predicting or projecting the 12-hour pressure change ahead for two 12-hour periods. It is found that the maximum 24-hour change will be centered half way between the successive centers of 12-hour pressure change and equal to $\frac{3}{2}$ of either. If the change in 12 and 24 hours is known, it is a simple matter to construct the actual pressure maps.

It may be true in Europe that the pressure changes associated with the westerly type move appreciably at the rate of 700 km. in 12 hours in winter, and 500 km. in summer, but this is certainly not true in the United States. Further, the shape of the change area and the magnitude of the changes undergo large variations in the United States. As the assumptions on which the system is based are not valid in the United States, they cannot of course be employed here.

Although knowledge of the pressure distribution is most important, it is by no means the complete story, and the French have extended their studies in attempting to associate the occurrence of precipitation, cloudiness and changes of temperature, etc., with the pressure changes.

Cloud systems.—Under the title of Cloud Systems [10] the French have published three volumes. The first volume contains the text, the second weather charts showing the movements of the cloud systems and precipitation and their relation with the isobars and pressure change areas, and the third volume is made up of photographs. The pressure changes, "noyaux," particularly the falls of pressure in 12 hours, are active moving systems which are distinguished from the low pressure areas when the changes are slow especially in the latter. The "noyaux" are shown by isallobars which have been long used by the French forecast service. They may show, as closed systems, the small secondaries which only appear as sinuosities on the isobaric chart and which are an important feature of the weather in France. According to this system each "noyau" or pressure fall area has its characteristic cloud system and it is found that the "noyau" and the main body of the cloud system are carried along in the general currents prevailing between the 2 and 5 kilometer levels. The charts illustrate the movements and persistence of areas of cloud and rain and also their final break-up and dissolution. The absence of cases of development may be attributed to the fact that most of the systems over western Europe develop over the Atlantic. The cloud systems are grouped into "systèmes dépressionaires," "systèmes d'alto-cumulus" and "systèmes orageux." The first includes an area of

continuous precipitation and is preceded by the well-known cyclonic cloud sequence. The front and body of the system are regarded as a consistent whole. The "systèmes d'alto-cumulus" are characterized by predominance of alto-cumulus cloud in the body of the system as well as on the flanks, and by the relative scarcity of cirrus forms. The "systemes orageux" are shallow, irregular, and slow moving, the clouds and precipitation being irregularly distributed. The most typical cases are the shallow thundery depressions of summer, which have characteristic cloud forms at various levels.

D. The Norwegian Method and the Structure of LOWS.—In earlier years it was thought that the LOWS were vortices and that the cyclonic circulation shown at the surface persisted up to considerable elevations. It was not until the advent of numerous kite observations that these older ideas of the structure of LOWS were definitely discarded.

Cyclone model.—The Norwegians under Bjerknes have presented a model of the structure of the cyclone that conforms in most ways to what is actually observed in nature. It involves two essential "fronts," the "warm front" along which cold air gives place to warm air, and the "cold front" along which warm air is replaced by cold. The fronts, especially the cold front, were familiar to many before the Bjerknes theory was promulgated, the earliest reference perhaps being published by William Blasius, Philadelphia, 1875.

According to the Bjerknes theory,[20] a cyclone consists of two essentially different air masses, see Figure 96, a warm current of south or southwest winds in the region of the cyclone and its trough, with cold air at the rear and also at its front. These air masses are separated by fairly distinct boundary surfaces which pass through the center of the cyclone. Such surfaces are inclined in the vertical always toward the cold side at an angle or slope of the order of 1 to 150. The middle portion of the figure shows the plan of the surface air currents about a low. At the center of the low the warm and cold fronts (broken lines in figure) meet. The upper portion of the figure shows a vertical east-west section through the cyclone north of the center. Here the air at the surface is moving from an easterly or northerly quarter and is relatively cold as we frequently find in the northern quadrants of a LOW. Aloft the wind is from the south or southwest, and being of tropical origin is relatively warm. It came originally from lower levels and lower latitudes, having undergone a forced ascent over the cold wedge of surface air, with the result that there is considerable cloudiness and precipitation due to the cooling of air through ascent and expansion. The lower cut shows another vertical east-west section through the cyclone south of the center. At the left are seen the cold polar surface winds from the west and northwest

underrunning the warm tropical south and southwest winds, which latter persist aloft after the wind has changed to westerly at the surface. This underrunning results in a forced ascent of the warm air from the south with resulting cloudiness and heavy, and, as a rule, brief showers. This front is called the "cold front," and is identical with the squall line or trough. At the right we have the relatively warm southwest winds over-

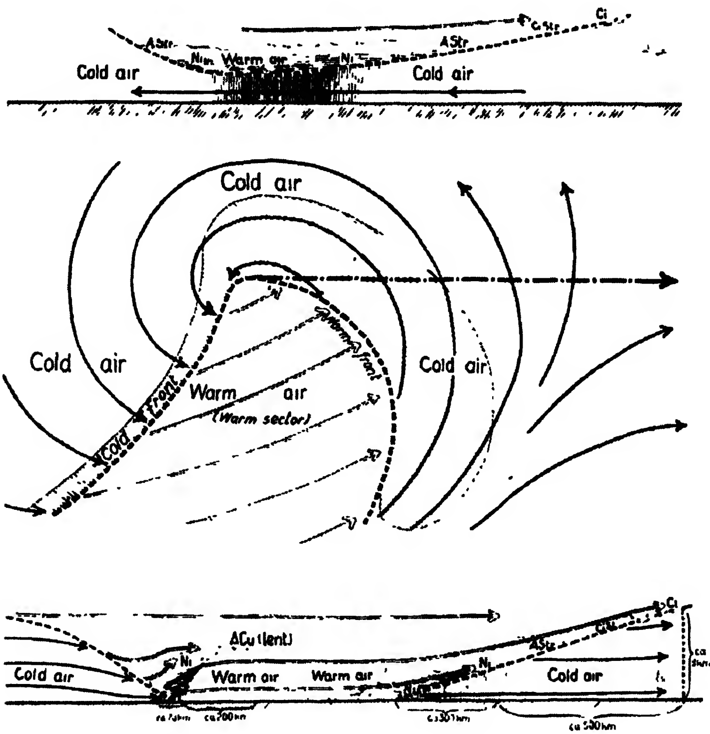

Fig. 96.—Bjerknes' cyclone model.

running the colder air to the east, giving widespread cloudiness and precipitation. In the central figure, the broken line trending east and southeast from the low center shows the position of the "warm front," while the "cold front" extends southwest from the center of lowest pressure. In the United States between the warm and cold fronts, at and near which cloudiness and precipitation are in evidence, there is a region of warm air and relatively clear sky and sunshine, although cumulus clouds are to be expected because of conditions favorable to convection. It will be noted that a narrow band of showers occurs along and immediately behind the cold front due to the raising of the warm south and southwest

current by the cold west and northwest current underrunning it, thereby adiabatically lowering its temperature below the dewpoint.

The vertical structure of the front is often diagrammed as shown by the line *AB*, Figure 97. But it seems more likely and is borne out in some cases by observations that such fronts have a form more like that shown by the line *CDE*. This is due to the fact that ground friction retards the movement of air near the surface with the result that the air aloft, say at 1,500 or 2,000 meters, runs ahead of the surface air. In such cases there is bound to be a certain amount of warm surface air trapped below the nose, *D*, of the colder air aloft with resultant instability and brief showers and possibly thunderstorms. On and immediately ahead of the warm front, however, the area over which rain is falling is relatively broad, the rain being due to the forcing of the warm south or

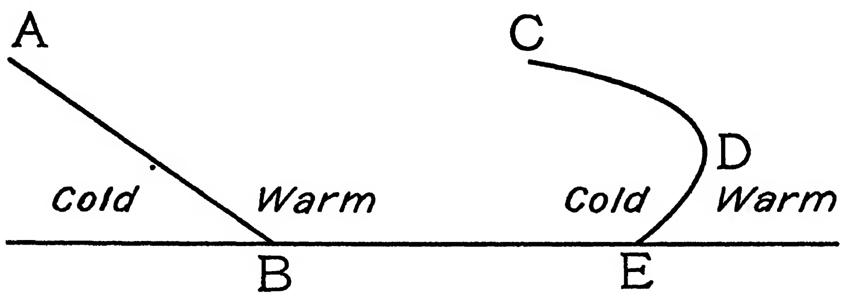

FIG. 97.—Cold front in vertical section.

southwest current up over the cold and dense currents from the east and northeast, thereby producing adiabatic cooling. What may be a cold front today may become a warm front tomorrow, and such we find occurs in certain cases. Fronts are said to be stationary when neither warm nor cold air is gaining ground along them.

According to the Bjerknes' ideas, LOWS have a warm sector due to currents of tropical air during the early part of their history. It has been noted, however, that occasionally we have the development of LOWS in polar air and that such depressions do not have a warm sector. Douglas [21] calls attention to a formation of this kind off western Ireland on November 10, 1926, followed by another about a week later.

In considering the structure of LOWS, attention should be called to the fact that the low pressure in the lower levels is frequently accompanied by relatively high pressure in the high levels. An example cited by Douglas illustrates this condition. He states that in comparing the observations on December 22 and 23, 1925, it is found that while surface pressure was 30 mb. (.90 inch) lower on the former date, the pres-

sure at 9 km. was 9 mb. (.27 inch) higher and that the base of the stratosphere was 2.7 km. higher. These ideas have been developed and supported by German writers.[22]

The structure of LOWS has been fairly well covered in the section immediately preceding but it will be well to add a word about the winds between 2,000 and 4,000 meters in LOWS of average or greater velocity of movement in the United States. Prior to the passage of a LOW the prevailing winds at these levels are between west and northwest in middle and northern districts. Concomitant or possibly slightly preceding the passage of a LOW as shown by surface isobars, the winds shift to west-southwest or southwest, and occasionally to south. As the LOW moves away the winds shift back again to between west and northwest.

Let us now consider winds between 2,000 and 4,000 meters, in average or fast moving anticyclones. After the passage of a LOW, the winds generally shift to west and northwest as the HIGH approaches, continuing in these directions until shortly after the center of the HIGH has passed, when they again shift to west.

In the case of slow moving HIGHS and LOWS, a somewhat different situation prevails in the 2,000 to 4,000 meter level. In conditions approaching stagnation, winds in the eastern half of the LOW will often be between south-southeast and south-southwest, while on the eastern side of the following high, north or north-northwest. winds will obtain. Such opposition of winds in a relatively small horizontal distance indicates a continuation of the stagnation or slow movement. In winter LOWS, winds with large southerly components occasionally attain altitudes of 4 to 6 kilometers, while in the summer HIGHS a few cases have been noted of winds of large northerly components up to 6 and 8 kilometers. It may be said that winter LOWS are deeper than those of summer and winter HIGHS shallower than summer HIGHS in their wind structure.

Life history of LOWS.—Formation: Cyclones frequently occur in families on the line and surface of discontinuity separating cold polar air from warm tropical air. The point where the development takes place, according to the Bjerknes' ideas, is where a wave develops in the line or surface of discontinuity, generally to the southwest of the anticyclone.

Occlusion: As the cold front advances more rapidly than the warm front, they gradually get closer and closer together, and finally meet when the south and southwest current constituting the warm sector gives place to the cold polar air at the surface first, and gradually to greater heights until the supply of warm, tropical air has been shut off completely. Such a process is called the occlusion of the low. Following an

occlusion, the low generally decreases in intensity owing to the fact that the warm moist southerly current which constitutes one source of its energy has been shut off.

Regeneration: Regeneration or re-development is said to take place when a LOW once formed decreases and later increases in intensity. An extremely interesting case of this kind is described in the Monthly Weather Review.[23]

Application of the different models to forecasting.—That the application of these cyclone models to forecasting throughout the north temperate zone at least has been generally accepted is evidenced by a considerable literature on the subject of the application of the front theory to weather prediction including detailed studies of many special cases. Although there yet remain some cases of precipitation occurrence to be accounted for, it is true that the application of the front ideas has afforded a quite satisfactory explanation of many cases heretofore shrouded in doubt. Only a few studies regarding conditions in the United States have been published. The first of this nature by J. Bjerknes and M. A. Giblett, concerned the development of a disturbance of considerable intensity off the south Atlantic coast in October, 1923.[23] Later papers covered conditions over the land, particularly with reference to precipitation.[24, 25]

It will not be possible to repeat the points brought out in these papers in this publication and the reader is referred to the original papers cited.

Neither will it be possible to bring out detailed applications to local cases, but only to indicate some of the broader principles involved. Regarding the movement of LOWS it has been pointed out by European writers that a LOW with a well-developed warm sector is moving with the same speed and in the same direction as the winds in the warm sector at an elevation, say, of 500 m. It has further been pointed out and again by European writers that when a LOW is occluded, *i. e.,* no longer has a warm sector, it will move in the direction and with the speed of the prevailing wind currents at approximately 2,000 m. Gregg [26] has called attention to the fact that the average 24-hour rate of movement of LOWS in winter corresponds quite closely to the average wind speed at 3,000 m. and the average rate in summer to that at 4,000 m. An unpublished study by Welby R. Stevens and the present author would seem to indicate a somewhat lower level, at about 2,000 m. It is found, however, that, while on the average the rule holds quite well, there is a great lack of agreement in individual cases. The lack of agreement was so marked in one series of maps that we modified our approach to the problem by attempting to find the level in which the speed of movement at the current 8 a. m. observation corresponded with the subsequent 24-hour move-

ment of the LOW in speed or direction, but without developing any general conclusions. The 24-hour movement of the LOW was also compared with the speed and direction at different levels observed at the middle of the 24-hour period, but likewise without success. The results were so discordant that we would hesitate to attempt to lay down any specific rules for low movement based on winds aloft.

Some relations between the movement of HIGHS and the wind speeds at different levels have been pointed out by Mitchell.[27]

Exner has given figures bearing on the angle of slope of the surface of discontinuity as a function of the surface temperature difference and velocity difference. So far as known, only a limited number of applications of these relations have been essayed in the United States, but they have proved to be in general harmony with conditions as shown by pilot balloons and kites. In estimating the speed of movement in the free air at an elevation of 500 meters, a rough approximation is furnished by taking twice the speed of the surface winds. In comparing temperatures in elevated regions in connection with the location of fronts, it is necessary to reduce the temperatures to the same plane, which can be done roughly by allowing 1° C. for every 100 meters difference in elevation.

The occlusion of a LOW is generally reflected in rising pressure at the surface throughout the region of the trough. The advance of a warm front at the surface is generally preceded by falling pressure.

Modifications to be applied in the United States.—In this country the Pacific States enjoying a west coast climate have in many ways characteristics common to western Europe. From the Rocky Mountains eastward, however, the climate is mainly continental, and thereby differs materially from the climate of western Europe, with the result that the rules developed by European writers in the application of the polar front theory will have to be modified accordingly. While the modifications in the application of the polar front theory west of the Rockies may not be considerable, yet the difference in the trend of the coast line, north-northwest and south-southeast in the United States and south-southwest and north-northeast in Europe, together with the gradual upslope of Europe from west to east, compared to the more precipitate ascent to the Rocky Mountain Range in our western states warrants us in considering these peculiarities very carefully. In the Plains region, we have considerable down-slope effect for the prevailing winds, complicated further by the fact that moisture-bearing winds are south and southeast. Between the Mississippi River and the Appalachians again we have up-slope effect for the prevailing westerlies with moisture-bearing winds between southeast and southwest; in the Atlantic States we have the down-slope effect for the prevailing westerlies, but up-slope effect for the moist winds

between northeast and southeast. No doubt in the upper Mississippi Valley, the northern Plains States, the Lake Region, and the Ohio and Mississippi Valley, we have the greater amounts of precipitation caused by moist winds directly from the Gulf, but frequent showers occur owing to moisture evaporated over the land from precipitation which has previously fallen. It will no doubt have been noted that we have precipitation, generally light, in these regions a number of times in the year when we do not have surface or upper winds from the Gulf either directly or indirectly. Further, after a drought in this area, the first consequential rains almost always appear over the West Gulf States and spread northward, indicating that the moisture comes originally from the Gulf.

It has been remarked by European writers that the great majority of depressions with which they have to deal are of the "occluded" type. In the United States, however, between 40 and 50 per cent of our lows develop over the continent, or waters contiguous thereto.

Air mass analysis.—A great many studies have been made in Europe, particularly by the Scandinavians [28] in identifying air currents in the different levels over a station as being composed of tropical or polar air, the polar air being generally thought of as dry and cold and the tropical air as moist and warm. These classes have been each subdivided into continental and ocean classes, as of course there are material differences between the two. Great assistance in forecasting has been gained through these studies in western Europe where the climate is typically maritime. It is doubted by those who have considered the situation that the same advantageous results can be obtained from air mass analysis over areas where predominatingly continental climate obtains. However, such analysis would undoubtedly be profitable along our western coasts. Only recently Willett [29] has published the results of his studies in Europe and their application in the United States, especially as concerns fog in the eastern United States. In his article, it is pointed out that fog most frequently occurs in old polar air that has passed over a water area, showing the importance of a careful analysis, even if made in a general way, when upper air temperatures and humidities are not available.

3. LOCAL FORECASTING

A. Topographic Influences.—In general forecasting, as well as in local forecasting, the effects of topography have an important influence. This is true in forecasting winds, both as to direction and force. The same is true with regard to temperature, fog, cloud, and other elements. The raising or lowering of an air mass has an immediate effect on the pressure and at the same time on its temperature and as a corollary the amount of moisture that may be present. According to theory an increase

of 100 meters in elevation results in a decrease of 1° C. in the temperature for air in the so-called dry state. Pressure decreases roughly at the rate of .10 inch per hundred feet at least in the lower levels. It will readily be apparent that the temperature of any air mass that is forced to mount up-slope will decrease and, if this up-slope motion is continued long enough, the temperature will reach the saturation point and condensation will begin. As a specific case, let us consider a southeast current on the south Atlantic Coast. The southern Appalachians are roughly 2,000 feet (700 m.) in elevation. If an air mass at the coast should have a temperature of 20° C. (68.0° F.) and a relative humidity of 70 per cent, an increase of elevation of 700 m. would decrease its temperature by approximately 7° C. (12.6° F.) to 13° C. (55.4° F.) with a resulting increase in relative humidity to approximately 100 per cent—enough under proper conditions to cause precipitation along the upper slopes of the mountains. Investigations bringing out the need for considering topography in local forecasting in Scandinavia have been made by Bjerknes.[30]

B. Summer Convection.—One of the most common examples of condensation by cooling brought about by ascent and decreased pressure is the formation of cumulus clouds of the summer season. Here, due to local heating, air is forced to rise, its temperature being decreased adiabatically at the rate of approximately 1° C. per hundred meters of ascent, with the result that, if the air contains a moderate amount of moisture, the saturation temperature is reached at an elevation of roughly 1,500 meters, being manifest in the formation of a cumulus cloud. The height of the cumulus base remains nearly stationary but rises slightly as the maximum temperature of the day is approached. A rough measure of the height of the base of the summer cumulus has been given by Espy[31] and Clayton[32] as follows: Dry bulb − dew point ÷ 0.78 equals height of cloud base, in 100s of meters. The divisor 0.78 is employed when C. degrees are used and 1.4 when F. degrees are used. A rough idea of the time cumulus may begin to form is also to be obtained from the morning (8 a. m.) humidity, the lower the humidity the later the clouds will form. Better information can, of course, be obtained from a morning airplane flight giving temperatures and humidities aloft. The higher the extent of the morning inversion, the later will the cumulus form and the higher will be the base of the cloud. Winter convection is important in its effects on the wind speed, especially at the surface. In the region to the southwest of a low center and east of a HIGH, it is frequently noticed that winds die down toward and during the night (unless there is a large surface temperature gradient to the westward) and increase again after the sun has risen. This effect is due to the fact that relatively high veloci-

ties aloft exist both day and night and the surface winds that have died down during the night increase as soon as convection is sufficiently established to cause a mixing of the lower with the upper currents, with the result that the higher velocities aloft are brought down with some modification to the surface.

C. Fog and Haze.—Fog is one of the most important obstacles with which navigation has to contend, as it occurs both over water and over land. Water fogs are classed under the advection type, while those over the land involve for the most part radiation. Fog constitutes one of the precipitation phases which is confined to the lower part of the troposphere and is one in which nuclei of condensation play a most important part. It is in these lower air strata, particularly in industrial regions, that local pollution of the atmosphere produces such a large number of condensation nuclei that they exercise a marked control of fog frequency, density and duration. Willett [29] outlines the main classes of fog as follows: 1) Advective type, 2) radiation type, and 3) type characteristic of maritime transitional air when cooled over land. The first class is subdivided, under the heads of (a) type due to transport of warm air over a cold surface and (b) type due to the transport of cold air over a warm water surface, the (a) type producing more widespread and deeper fogs than the (b) class. The subdivisions of the second class are (a) ground fog characterized by a surface inversation, (b) high fog characterized by an upper inversion and (c) inversion haze.

In advection fogs it is important to have a moderately high absolute humidity which may be increased by evaporation, or it may be decreased by the deposit of dew and by condensation in the form of rain or snow. The vertical temperature gradient should be small, being subject to increase by relative heating in the lower levels or relative cooling in upper levels; and vice versa, it may be decreased by relative cooling in the lower levels or relative heating in upper levels. The horizontal pressure gradient as it controls the northward flow of tropical air or the southward flow of polar air is also of importance.

Radiation fogs are favored by almost complete stagnation of air. Therefore, the forecasting of fogs of this type is dependent upon a consideration of the modifying influences of local conditions upon the air mass already present and at rest, rather than on a consideration of the modifying influence upon a distant air mass of the route traveled by that air mass. In this case of the stagnating air mass the same properties of the mass are significant as for the formation of advection fog, namely: absolute humidity, and vertical temperature and horizontal pressure distribution. But in this type of situation the lightness of the wind velocities is such as to permit of a sufficient local concentration

of active nuclei of condensation, so that this becomes a fourth property of the air mass significant for fog formation. The observation is made that there cannot result a general fog formation of any appreciable depth in cold air over a warm damp land surface any more than there can result a general sea fog over warm water, because the formation of a surface inversion is equally impossible. The actual dissipating effect of a warm water surface is shown by the often observed thinning of early morning radiation fogs over lakes and rivers, when there is no marked valley gravitational effect to counteract it. This principle is so well known to German aviators that it is a regular practice in the case of low radiation fog to change their route so as to follow the course of a river.

Ground fog is densest near the ground and diminishes in density with increasing elevation. It is characterized by an inversion whose base is at the ground and upper edge at or above the top of the fog. It is in the vicinity of large cities that this type of fog appears in its worst form.

The phenomena sometimes called high fog, but which in reality is cloud of stratus or strato-cumulus type, is characterized by a marked inversion above the ground (from 200 to 2,000 meters) just as the ground fog is characterized by a surface inversion. The greatest density of the fog is at the base of the inversion, just as the greatest density of the ground fog is at the ground.

Any general discussion of fog forecasting must be based on theoretical considerations rather than on principles applied in practice, for the simple reason that there is no developed system of practical fog forecasting. The dissipation of ground fog can usually be forecast with some certainty, but this is not true of the dissipation of the more stable types.

A precedent condition to the occurrence of radiation fogs is clear sky, which is associated with anticyclonic conditions. Further, they occur in air masses that have come from continental regions and that are thermally stable. Under such conditions, Taylor [33] has found that out of 70 cases of fogs of this kind, 62 occurred with winds at midnight less than 3.3 m. p. h., 5 cases with winds between 3.3 and 5.5 m. p. h., and 3 cases between 5.5 and 9.2 m. p. h This shows that a wind force greater than 3.3 m. p. h. even at midnight is frequently enough to prevent fog formation before morning. It has further been shown that in the complete absence of wind sufficient cooling under most conditions takes place up to an elevation of only 3 or 4 feet, probably attended by the formation of dew. Some slight air movement is necessary to dynamic turbulence which, if slight, does not produce sufficiently rapid dispersion to efface the ground cooling, but rather makes its influence felt throughout the lower layers instead of only throughout the ground layer. In this case a much greater mass of air is cooled. However, an appreciable air move-

ment increases the dispersion of the ground cooling, and the relative warming of the lower air becomes such that fog formation is out of the question.

The general conditions under which we may expect ground fog are: 1) mostly clear skies; 2) light winds, not more than 6 m. p. h., but not absolutely calm; 3) thermal stability and early evening appearance of ground inversion; 4) sufficient absolute humidity indicated by early evening deposit of dew. Wind direction favorable to carrying pollution to the location under consideration is not an essential but is important in some cases.

A paper by Entwistle [34] stresses the importance of topographic influences in fog occurrence. He also calls attention to the use of captive

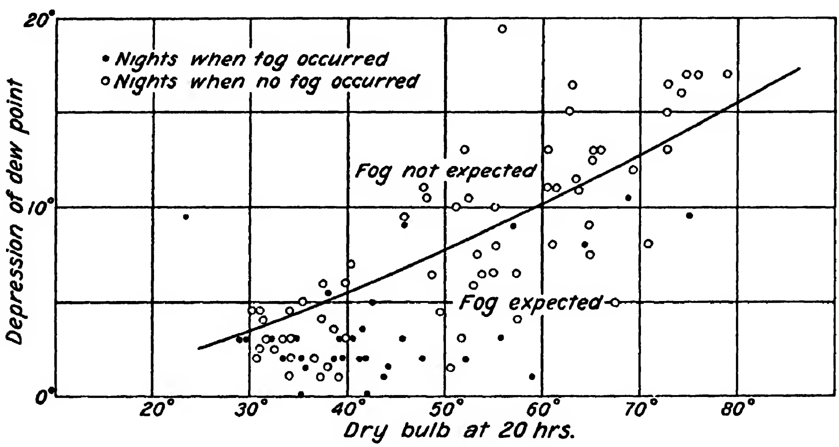

FIG. 98.—Taylor's fog graph.

balloons equipped with an instrument for recording height above ground, relative humidity and temperature, to determine the thickness of the fog, principally from the temperature record.

Any discussion of radiation fog would be incomplete without a reference to Taylor's graph shown in Figure 98. He simply took for a given station all evenings on which the formation of ground fog seemed possible, that is, with clear skies and light wind, and represented the conditions in a field of coordinates where the ordinates represented the depression of the dew point and the abscissæ the temperature at 8 p. m. The conditions each evening when fog occurred were represented by a dot in the field, a circle being used to indicate when no fog formation followed later in the night. Taylor found that a smooth curve could be drawn through such a field of points so that nearly all the points above the curve represented conditions which were followed by nights free of fog.

Some notes on fog distribution and forecasting were given by Frankenfield in an address before the Geophysical Union.[35]

D. Line Squalls.—While line squalls should be primarily considered general phenomena, they should certainly be mentioned as local phenomena as well, since aviation, including both lighter than air and heavier than air, is affected to a great extent by them. Therefore, for the meteorologist at the local air station, they constitute an important local problem.

The line squall is a cold front phenomena which occurs in conection with well-developed trough formations. The most exhaustive reference on the subject is to be found in a paper by M. A. Giblett,[36] which deals more particularly with such phenomena as they occur in Europe. In an article by the present writer which is published on the back of the Pilot Chart, issued by the Navy Department, for the North Atlantic Ocean for March, 1929, phenomena of this kind occurring in the United States are discussed. The line squall is characterized by a sudden shift of wind attended by a squall ranging from 15 to 20 m. p. h. up to as much as 100 m. p. h. Another essential characteristic is a sharp fall in temperature; in fact, this is the physically important element. The squall front may extend for several hundred miles in especially well-developed cases but is frequently much less. The speed of advance has a wide range, the most usual limits being from 25 to 35 m. p. h., but in some cases reaching 50 or even 60 m. p. h. These phenomena, including all degrees of intensity, are quite frequent; but for every severe one there are many of slight or only moderate intensity. They may occur at relatively short intervals of a day or two, but more frequently a period of several weeks may elapse which will be free from such phenomena. The violent motions peculiar to the line squall and the attendant roll cloud are limited in most cases to the first 6,000 feet above surface, but above this the usual intense vertical currents of thunderstorm clouds may occur up to 15,000 or 20,000 feet. The vertical currents in these well-developed thunderstorms often reach enormous velocities, as judged from the size of hailstones, in some cases exceeding 2½ inches in diameter. It is known from experiments that velocities of approximately 80 m. p. h. are required to sustain stones of that size. The region of line squall occurrence as related to the HIGHS and LOWS is in the trough of the LOW, where warm winds from a southerly quarter are replaced by cold winds from a westerly or northerly quarter. The shift at the surface is generally preceded by a typical line squall cloud or "Roll" cloud, an illustration of which is given in Figure 99, indicating that winds at an elevation of 1,500 to 2,000 meters (4,500 to 6,000 feet) have shifted to a westerly or northerly direction, while the winds at the surface remain for a short

period still from a southerly quarter. No complete survey of any one case at all heights has ever been made, but there are many fragmentary observations available of various phases of the different cases. Piecing these together and combining them with non-instrumental observations and certain theoretical considerations, a picture of the essential features in the immediate neighborhood of the squall line may be obtained, as illustrated in Figure 100. In this diagram the horizontal and vertical scales are the same, so that a correct idea of the proportions is at once given. It should be borne in mind that the line squall is essentially a cold

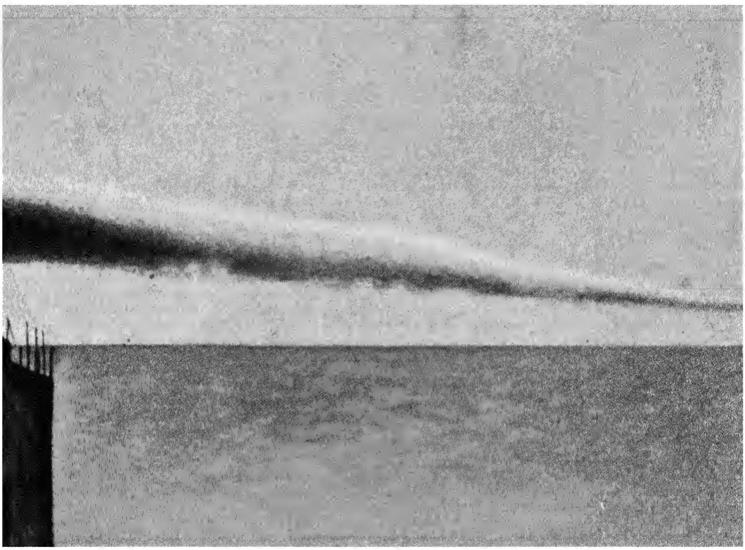

FIG. 99.—Line-squall cloud. (Courtesy of Royal Aeronautical Society.)

front phenomena and that, unless cold air replaced warm air and also overran it, line squalls would not occur. In the diagram the hatched area is a section of the cold wedge which is advancing to the right. Owing to surface friction the cold air is found further advanced above the surface than at the ground itself, and the upper limit, AB, of the cold air, instead of continuing until it cuts the surface, is doubled back at the point B and meets the surface of the ground at the point C. This latter point is to be considered as marking the cold front at the ground. The moment when the barometer begins to rise at the ground station is when the point B comes overhead and, as the denser air above the station increases in thickness, the barometer continues to rise rapidly. It is on the passage of the point E that the squall commences and continues until a little

after the passage of the point *C*, when the wind shifts to a westerly quarter. The letter *F* in the figure indicates the position of the line squall cloud. The average distance between the cloud and the wind shift at the surface is about 4 miles. The time interval separating the two phenomena depends on the speed of movement of the whole system. If we assume a rate of movement of 30 m. p. h., the time interval would be eight minutes.

E. Visibility.—Fog and haze constitute the most important factors affecting visibility. Of haze[29] there are two types, the ground inversion and the high inversion. The principal source of the haze is dust or smoke from large cities, forest fires, and the surface of the ground when dry. The density of ground haze is greatest near the ground. Owing to the

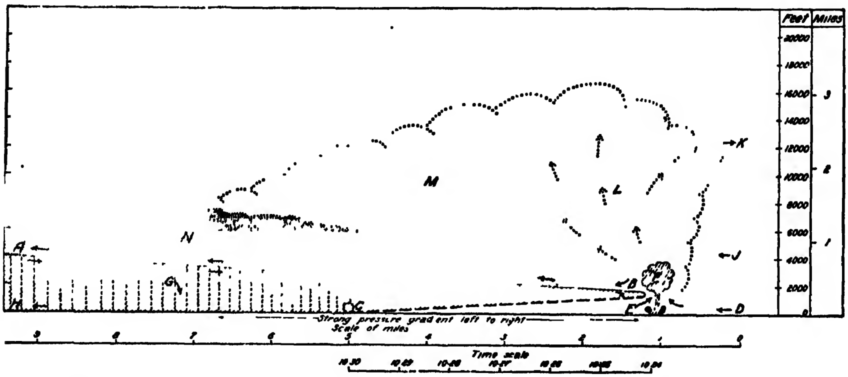

Fig. 100.—Enlarged scale vertical section of line squall. (Courtesy of Royal Aeronautical Society.)

almost complete calm which characterizes a ground inversion, it is of necessity extremely local. It can form only in the immediate vicinity of the polluting source. High haze is of more importance, both practically and theoretically, than ground haze. The most characteristic thing about high haze is its stratification in distinct and homogeneous layers of the atmosphere.

In the vicinity of the source of pollution, the density of the inversion smoke haze is directly proportional to the amount of the inversion and inversely to the wind velocity. Farther from the source, the principal factor is the direction of the wind related to the source. In summer, the continual convectional heating does not permit of high inversions and consequently there is little high haze; but ground inversions and ground haze may occur in summer. The fact that objects above the haze may be quite distinctly visible to one on the ground, while at the same time the ground is entirely invisible to the observer above the haze, is due to the

efficiency of the haze layer as a reflecting and scattering medium. The forecasting of haze is easier than that of fog, for the haze obscurity is directly proportional to the concentration of the pollution and not dependent on the attainment of any saturation concentration, as is the case with fog before it becomes visible. The essential factor in forecasting ground haze is the ground inversion.

Other things being equal, the density of the haze is proportional to the amount of the inversion. Ground haze is very local in the region of the source of pollution. What wind there may be carries the haze with it, so that the expected direction and velocity of the wind must be considered in forecasting the haze distribution.

In the case of high inversion haze the principal factor is the high inversion which accompanies an extensive anticyclonic development, the same being true of high fog. As soon as such an inversion becomes marked, and this happens invariably in the settling of the polar air mass, the high inversion haze begins to appear. For this phenomenon the dryness of the air makes no difference. Since this is a persistent phenomenon, the high inversion haze may gradually increase in extent and density day after day until it becomes a general nuisance over a whole region. In forecasting the horizontal distribution of the more general haze, the primary factor to take into consideration is not the rather light wind in the lower strata below the base of the inversion, but the direction and velocity of the stronger winds in the upper part of the inversion layer and the base of the layer next above. The density will be greater if the winds are light but the haze will be more widespread if the winds are fresh. The upper level of the haze stratum is rather abrupt, just above the top of the inversion, and its elevation may be forecast accordingly.

The visibility scale employed in the United States and in Europe is given below:

HORIZONTAL VISIBILITY SCALE

0 =	Objects	not	visible	at	50	meters	(55	yards)
1	"	"	"	"	200	"	(220	")
2	"	"	"	"	500	"	(550	")
3	"	"	"	"	1,000	"	(1,100	")
4	"	"	"	"	2,000	"	(1¼	miles)
5	"	"	"	"	4,000	"	(2½	")
6	"	"	"	"	10,000	"	(6¼	")
7	"	"	"	"	20,000	"	(12½	")
8	"	"	"	"	50,000	"	(31	")
9	"	"	"	"	50,000 or more	"	(	)

F. MAXIMUM AND MINIMUM TEMPERATURE.—The forecasting of maximum temperatures is of considerable importance when the comfort

of the public is concerned, especially in the warm season. However, it may have far-reaching effects during the spring season in connection with the melting of the snow cover and the consequent effect on the stages of rivers. It has been found at the Washington office that the morning aeroplane flight furnishes quite a dependable basis for maximum temperature forecasts for the afternoon. If the upper air temperatures are relatively high and the morning inversion shallow, there will be a rapid rise

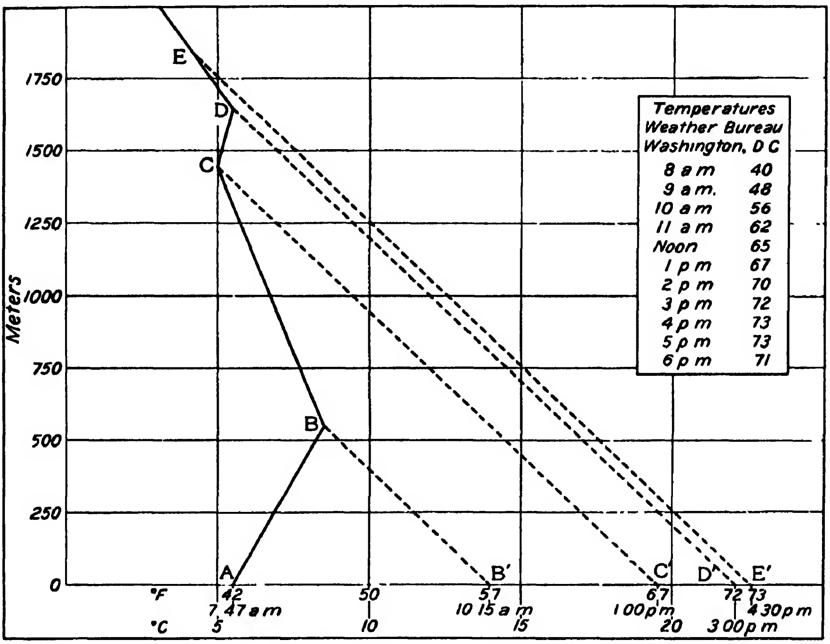

FIG. 101.—Temperature graph, March 13, 1930, Naval Air Station, Anacostia, D. C., showing relation between vertical temperature distribution and occurrence of maximum temperature at the surface.

until the potential temperature of the upper part of the inversion is reached, when the rise will be checked.

It is also possible in most cases to estimate very closely when the rapid rise will cease and the gradual rise begin. To illustrate, see Figure 101, which shows the vertical temperature distribution as obtained from an airplane flight made at 7:47 a. m., March 13, 1930, at the Naval Air Station at Anacostia, D. C. It has been noted that, if the point *D* at the upper limit of the inversion is projected downward along the adiabatic line to the base of the figure, it will approximate quite closely, or be a little below, the maximum temperature for the afternoon. In the figure the projection of the point *D* to the point *D'* on the base line corresponds

to 22° C. (71.6° F.), whereas the maximum temperature was 22.8° C. (73.° F.). It will be further noted that the rise of 15° in temperature from 42° to 57° covered a period of two hours and 28 minutes, or at the rate of 10 minutes per degree; the rise of 10° from 10:15 to 1:00 p. m., was at the rate of 16.5 minutes per degree; the rise of 5° between 1:00 p. m. and 3:00 p. m. was at the rate of 24 minutes per degree; and the rise of 1° between 3:00 and 4:30 occupied about 90 minutes. From this it will be seen that the rise becomes progressively slower and slower.

The prediction of minimum temperatures is of prime importance in the orchards of the west, where resort is had to orchard heating. The principal basis for local forecasts of this kind is the so-called hygrometric formula developed by Donnel,[37] Nichols, Smith,[38] Young, Ellison and others.

A recent article [39] summarizes the development of the latest formulae. Formulae of this kind are applicable only to the so-called radiation nights when the sky is clear and the air relatively quiet. Improvements and refinements have been made in this work by considering barometric gradient, since small differences of wind velocity, even in comparatively light winds may affect the results by as much as 4° to 6° F., which, when temperatures are approaching a critical degree, are of the utmost importance.

G. Frost.—Dew occurs on clear nights when the temperature of the ground and objects on and near the ground falls to or below the dewpoint. In such cases, when the dewpoint is 32° F. or lower, frost occurs. Conditions fulfilling the above requirements occur under anticyclonic distributions of pressure, when clear skies and conditions most favorable to rapid loss of heat by radiation are experienced. When the air is quiet, air drainage due to topography is an important consideration. Sometimes frost will occur at a place separated by only a hundred feet from another place where no frost occurs. It is well known that the earliest frosts occur on valley floors rather than on the slopes. The practical importance of considerations of this kind is shown by the distribution of vineyards along the banks of the Rhine, the vineyards being located on the slopes. That there exist such thermal differences has been learned through bitter experience. In the region of western North Carolina careful studies have been made by Cox,[40] showing the existence and delimiting the extent of such thermal belts.

Plants on clear nights when the net loss of heat by radiation is large lose heat much more rapidly than does the air. In the cranberry marshes of Wisconsin it was found that minimum thermometers exposed in the open 5 inches above the ground and touching the tops of the vines averaged for an entire season, 2.8 ° below those at the same elevation above

the ground in shelters nearby. On clear quiet nights the difference was largest, the greatest being 9.9°, but there was little difference on cloudy and windy nights. Both green vegetation and dark soil are excellent radiators. Frost is more likely to form when there is a moderate amount of moisture present in the air. In dry air the temperature may fall below the freezing point without any evidence of frost formation.

Frost often occurs on still nights on valley floors while points higher on the slopes escape, the temperature sometimes differing as much as 10 to 15° and even 20° over limited areas. On windy nights the temperature does not differ materially at neighboring points in hilly country, as the warmer air from above is being continually mixed with the cooler air near the surface. Damp soil favors the occurrence of frost but a very wet soil has the opposite effect. Frost infrequently occurs following heavy rains but a light rain on an overcast, windy day is favorable. Then the initial temperature in the evening is low and the soil is cold because of the evaporation of moisture from the surface and the fact that there was no heating of the soil during the day. When ground is moderately wet and frost occurs, it is generally light.

II. Storm Winds.—When storm winds are once in evidence on the map under the influence of a well-developed LOW and sharp pressure gradients, the situation does not provide so difficult a problem for the forecaster as it does in the case of a new and sudden development of a LOW, often in the shape of a secondary, or a rapid increase in intensity that requires the utmost care in handling. The two most common forms of sudden development described by Bowie,[12] are the "swinging trough" type and the "neck of high breaking off" type. In the first, a low center of moderate intensity may be advancing eastward along the northern border attended by a "V"-shaped trough and the secondary develops over the Gulf or South Atlantic States. The best evidence so far found as to the formation of this type is a cyclonic wind circulation attended by a localized 12-hour fall of pressure in the southern end of the trough. If both occur the secondary will certainly develop. The second type is characterized by a ridge of high pressure extending eastward for a considerable distance from the primary HIGH which is generally located over the Rocky Mountains or the eastern Slope. The eastward extension of the HIGH breaks off or separates from the primary and passes to the east or northeast as an independent HIGH though small in comparison to the former. The separation of the eastward extension is generally indicated by the pressure change chart on which will probably be observed a 12-hour pressure rise over Tennessee or the Ohio Valley, separated from the main pressure rise to the westward. As the easterly or secondary HIGH moves east a LOW develops in the neighborhood of Florida, and

passes up the Atlantic coast. This type is found in a number of positions, with the main high and the extension in the same relative positions, but oriented differently.

One type of an increasing LOW may be mentioned. In this type the LOW lies south of the belt of relatively high pressure which connects centers of slightly higher pressure to the northwest and northeast, respectively. The LOW belongs to the oval type and the lowest pressure is in the southern end of the oval. Just as soon as the eastern HIGH increases its distance from the western HIGH, a marked pressure fall is shown north of the low center, and the center in the southern end of the oval moves rapidly northward with great increase in intensity. In many cases high winds accompany LOWS that deepen rapidly; in others the high winds are due to the marked barometric gradients between the LOW and the accompanying HIGH.

I. Sleet and Ice Storms.—Sleet is precipitation that occurs in the form of frozen or partly frozen rain, and is formed by rain falling from a relatively warm stratum of air into and through another air stratum that is sufficiently cold to freeze some or all of the rain drops. The term "sleet" is sometimes confused in its acceptation by the public with "ice storm" conditions. The latter term refers to precipitation that falls to the surface as rain, but that freezes as soon as it touches the surface. It often forms on telegraph and telephone wires, trees, etc. Sleet does not cling and causes little if any damage. It most frequently occurs with a surface temperature between 22° and 28°,[41] generally to be found in the southwestern periphery of an anticyclone, to the south or southwest of which is a low pressure area. In the terms of frontal meteorology, supported by upper air observations, sleet occurs just in front (ahead) of the surface warm front, attending cold winds between east and north at the surface, and with a warm south or southwest current aloft, generally at an elevation of 1,800 meters or above, which overruns the cold surface. The ice storm occurs under quite similar conditions as the sleet storm but nearer to the surface warm front line, the layer of surface cold air not being so thick as with sleet. The distribution and conditions favorable for the formation of sleet and ice storms have also been discussed by Frankenfield.[42]

J. Thunderstorms.—Thunderstorms may be divided into two main groups: 1) the convection or heat, and 2) the cyclonic thunderstorm. The first occurs sporadically due to local convection, attending flat barometric conditions or very weak barometric gradients and is typical of hot summer afternoons. The cyclonic type may be subdivided into 1) cold front, 2) warm front, and 3) overrunning type. The first type occurs along and immediately behind the cold front, those occurring

along the front being accompanied in most cases by the line squall. The ones that are noted behind the cold front are more common in the Plains States and lower Missouri Valley. The warm front type takes place ahead of the surface warm front. In both the cold front and the warm front types it is believed that what is termed "upper air convection" plays the initial rôle, that is, that warm, moist air with a lapse rate between the dry and the moist adiabatic is mechanically raised either by being forced up over the cold wedge ahead of the warm front or by being underrun by the cold wedge of the cold front, in both cases having potentially colder air above the warm layer, which results in instability. Attention has already been called, see Figure 97, to the effect of ground friction in conection with the cold front, with the result that some warm surface air is trapped under the cold nose aloft. In the cold front type, so far as has been observed, thunderstorms occur where surface temperature gradients are well marked, indicating underrunning at the surface and not the overrunning type to be discussed next, which occurs with slight or only moderate horizontal temperature gradients. The third type is the overrunning type, which occurs in the warm sector due to cold air in a westerly current overrunning the warm moist south or southwest winds observed in most cases at an elevation of 1,800 or 2,000 meters. Thunderstorms of this type occur in a line parallel to the cold front, sometimes as much as 200 or 300 miles ahead of the surface cold front. The warm front type is perhaps the hardest to forecast. The most frequent type is the convection type, but the overrunning cold front type is a good second, particularly in the lower Mississippi Valley and the Gulf States.

4. LONG RANGE FORECASTING

Reputable meteorologists and meteorological institutions have carefully refrained from making weather predictions for more than a few days or at most a week in advance, except in a very limited number of instances, the most important being the monsoon forecasts of India.

The problem of long range forecasting has been approached from numerous angles. The methods employed have differed materially but may be grouped under the following general heads:

1) Extraterrestrial influences
2) Periodicities and cycles
3) Relationships involving a definite time interval between conditions in different parts of the world (including snow and ice cover)
4) Ocean temperatures.

Before discussing what has been done under any of the above heads, it may not be inappropriate to call attention to the belief of some inves-

tigators, including the present writer, that it is unnecessary to invoke extraterrestrial influences, assuming of course a steady supply of heat from the sun, to account for most if not all of our daily and weekly weather changes and probably many of the seasonal and longer period variations as well. The distribution of land and water surfaces, the variations in the intensity and location of the major cold and warm ocean currents, the seasonal interchange of air between oceans and continents, the changing distribution of snow and ice fields, the rotation of the earth, together with the annual changes in the declination and distance of the sun, provide sufficiently varying influences to account for changes that take place in our weather.

A. Extraterrestrial Influences.—Considering briefly what influences from extraterrestrial sources might affect the atmosphere, it seems obvious that gravitational effects would be small and would affect large areas of the earth in much the same way, namely, by some general tidal action. Electro-magnetic effects as far as they affect weather seem to be ruled out, since the atmosphere is non-magnetic. The variations in solar radiation received by the earth are entirely too small to account for the marked weather changes which we in temperate and northern latitudes experience. Although considerable time has been devoted to investigations of this kind, especially by Clayton,[44] the results so far achieved, although interesting, have not been at all convincing.

Where to look for the manifestations of extraterrestrial influences except it be in the so-called "centers of action" of Teisserenc de Bort, or the *semi-permanent highs and lows* in the words of modern writers, is a highly speculative matter.

Under the head of extraterrestrial influences should also be included a mention of the planets and our satellite the moon. That the first of these may be eliminated so far as any definite relation with earth weather is concerned has the concurrence of opinion of the best investigators.[44] With regard to our satellite, it can be said that although a great deal of labor and time have been devoted to considering her phases, declination, distance and revolution of the nodes of her orbit, little if anything of value in the solution of the long-range problem has been gained. In discussing the so-called lunar effects, it will be necessary to refer to certain ideas that may also be logically mentioned under the head of "Cycles and Periodicities." These include the rotation of the line of nodes of the moon's path, which is accomplished in 18.6 years. Earlier investigations by Rawson and Lockyer seemed to indicate a period of 9 and 18 years in droughts and Hutchins also found something of this kind as related to South Africa.

With regard to the declination of the moon, Poincaré [45] found that the average barometric pressure on parallels of latitude around the whole globe as measured by the international maps published by the United States Weather Bureau, gives the following results: The pressure on latitude 40° minus that on latitude 10° is +1.88 mm. (.075 inch) when the moon is in extreme south and +4.82 mm. (.192 inch) when the moon is in extreme north declination. The normal difference is plus 3.35 mm. (.134 inch). This would seem to indicate that, when the moon is farther north, there is a slight accumulation of atmosphere in the northern hemisphere amounting to 1.47 mm. (.059 inch) on the parallel of 40°.

Sunspots have received more than an average share of attention. Köppen's earlier work was followed and supplemented by Mielke [46] who showed that a curve of sunspot numbers and the curve of earth temperatures follow or parallel each other in a general way, in the sense that the fewer the sun spots, the higher the temperatures with, however, puzzling discrepancies in a number of places. Although the discrepancies are marked, the agreement in the curves is sufficient to warrant the belief that there is some sort of a connection between sunspots and atmospheric temperatures. However, let us not suppose that the relationship is anywhere nearly sufficient to employ as a basis of prediction. The amplitudes of the temperature effects are extremely small as should be expected since the time unit employed is the year. Further, individual stations and individual years do not always conform to the rule. Walker [47] has pointed out that when relationships are very close a comparison of curves will show the fact, but in general the relationships between temperature and sunspots are far from close and only a correlation coefficient can give a definite idea of the connection between them. He computed correlation coefficients for 97 stations between sunspots and temperatures. The great majority of the coefficients were negative, showing agreement with Mielke's results. On the other hand, they were small in magnitude, the great majority being under 0.25.

B. Periodicities and Cycles.—An examination of almost any record of meteorological data will show, in portions of it as least, apparent repetitions of cycles. However, if a record of any considerable length is taken, a persistence of the cycle through the whole record will not be found. In 1889, for example, Hutchins [48] found that the droughts in South Africa had a cycle of 9 years, but shortly thereafter the phase was completely reversed and thus rendered the forecasts based on it useless.

The cycles that have received perhaps the most attention are the sunspot cycle of slightly over 11 years, the nodal cycle of the moon of 18.6 years and the Brückner cycle averaging about 35 years. Although the sunspot cycle averages slightly more than 11 years, the individual occur-

rences range roughly from 7 to 17 years. In the so-called Brückner cycle the range is from 30 to 50 years. To take care of such variations in length, Clough [49] developed the idea of periods, if we may call them such, of variable length, and for the sunspot and Brückner cycle has attempted to show the law by which they vary.

For elucidating periods and cycles, investigators have employed every conceivable means including graphical and statistical methods. Periodogram analysis developed by Schuster,[50] the well-known Fourier series of harmonic analysis, and accumulated sums [51, 52] have been used extensively, as well as smoothed values in graph form; [53] but all without developing results worthy of being used as a basis of prediction.

Investigators along these lines have produced some very interesting general relationships; but from the practical standpoint of predicting future occurrences in such a way as to be of commercial advantage, little has been accomplished and little has been claimed by conservative workers. In this connection Walker has pointed out, especially in relation to correlation coefficients, that there is no justification for issuing forecasts unless we have assurance of success in at least four cases out of five.

As an illustration of the periodogram analysis a work by Alter [54] may be cited. In this he has shown how the periodogram may be used in ascertaining periods in long rainfall records.

The harmonic analysis has been used in a number of investigations. One of the most interesting by Brunt [55] discusses the Greenwich temperature record carrying the investigation to 100 terms. Unfortunately the apparent periods that are developed by means of the Fourier series are artificial in that they are creatures depending largely on the length of record employed in the analysis. For example, if the length of the record is 48 years, the possible periods that develop are 24, 16, 12, $9\frac{3}{5}$, 8, $6\frac{6}{7}$, 6, $5\frac{1}{3}$, $4\frac{4}{5}$, 4 4/11, 4, etc., years; whereas, if a record of 60 years is used the periods become 30, 20, 15, 12, 10, $8\frac{4}{7}$, $7\frac{1}{2}$, $6\frac{2}{3}$, 6, 5 5/11, 5, etc., years.

Vercelli [56] originated a method based on the assumption that the weather is governed by a series of waves in the atmosphere which persist for a few weeks and then die away. He took a curve of the daily pressures for a period of about two months and analyzed the curve into its component waves. Then he projected these waves for a week into the future, after which he combined them into a single curve. This method met with an unusual amount of success for a time but with longer use did not live up to expectations and was abandoned.

In England forecasts published by the Daily Mail were based on 15 cycles, some of which were regarded as permanent and others evanes-

cent. These forecasts were made by Lord Dunboyne. A brief description of the method will be found in Nature (London), for January 29, 1927.

From a consideration of the results obtained, investigations of cycles have not produced encouraging results.

In connection with cycles it seems most logical to mention associations of preceding seasons with some future season. In this connection Wagner [57] showed a 16-year period in the temperature difference between the summer and that of the preceding winter at Vienna and other stations.

Another novel and quite interesting study was made by Griboiedoff [58] who associated winter conditions in Russia with temperatures of the following summer. He found that during the winter season each time the activity of the Siberian anticyclone is characterized by a pronounced positive departure in the region of the lower Obi and the Yenisei, it is followed by a period of atmospheric activity characterized by almost opposite departures. This phenomena may be repeated several times, constituting enormous oscillations of the atmosphere. He therefore separated the data for the season into these two groups and found that when the greatest oscillations in winter occur in the north the following summer will be warm; but when the greatest winter oscillations occur in the south, then the following summer will be cold. Attempts have been made to adapt this idea in the United States but results have not been satisfactory.

C. Relationship between Different Parts of the World.—Probably the most important and suggestive early work in this field was done in 1878 by Hoffmeyer who pointed out certain simultaneous relationships, for example between pressure in the north Atlantic Ocean and weather in western Europe. His work was followed by contributions from Blanford, Teisserenc de Bort, Hann, Meinardus and Pettersson. In 1897 impetus was given to this line of investigations by Hildebrandsson but as pointed out by another writer [59] the short period of ten years used in his investigations prevented him from reaching final conclusions. One of the most important things mentioned by him was the "see-saw" or opposition of pressures between Argentina and India. The Lockyers made this "see-saw" the basis of a classification of pressures over the world [60] according as they oscillated with India or with Cordoba in Argentina. In these investigations annual values were employed. Later, Walker [61] employed correlation coefficients, using seasonal (quarterly) values of pressure, temperature, precipitation and river floods at 32 centers distributed over the world.

The main conclusions reached by Walker [61] are that there are three big swings or surgings: 1) the north Atlantic oscillation of pressure between the Azores or Vienna on the one hand, and Iceland or Greenland on the

other; 2) the north Pacific oscillation between the ocean-high pressure belt and the winter depression near the Aleutian Islands; and 3) the southern oscillation mainly between the south Pacific Ocean and the land areas around the Indian Ocean. Of the three oscillations, the south Pacific is more far reaching than the others. From his tables of correlation coefficients, he found that Port Darwin (Australia) has no less than 76 significant relationships with other places, of which 32 are with subsequent seasons. He concluded therefore that pressure in the neighborhood of Port Darwin seems to exercise more control over other regions than any other world factor. He further points out that, although it may be some time before we learn the processes by which nature effects these enormous oscillations and the relationship found must in general be regarded as empirical, there is no reason why they should not be utilized when possible for seasonal forecasting. The facts of the southern oscillation have been utilized in predicting the rice crops of Japan [62] and the Java rainfall.[63] Tables have also been presented by Bliss [64] showing relations between Nile floods on the one hand and Dutch Harbor (Alaska), temperatures March-May; Samoa temperatures, December-May; and Port Darwin pressure March to May. The most hopeful results have been achieved in India [61] where a multiple correlation coefficient of .73 was obtained for the Peninsula rainfall, the regression equation being +0.44, South American pressure −0.29, Zanzibar rainfall −0.41, Java rainfall (October-February). Many other investigators have aided in developing relations similar to the ones mentioned above, among whom should be mentioned R. C. M. Mossman, C. E. P. Brooks, N. A. Hessling, Felix M. Exner, and others.

Another field involving relations between wind conditions over the oceans and weather has also received attention. C. E. P. Brooks [65] associated the seasonal distribution of pressure over the north Atlantic Ocean and western Europe with 1) the strength of the northeast trades 9 to 12 months before, 2) the strength of the southeast trades 12 months before, 3) the strength of the northwest winds between Newfoundland and southern Greenland which govern the strength of the Labrador current, and 4) the amount of ice in the east Greenland current and in the neighborhood of Iceland. Shaw [66] pointed out a remarkable similarity between annual variations of velocities at St. Helena and the annual variation of rainfall in southwestern England of the following year. A rather unique inverse relation has been brought to light by McEwen [67] between water temperatures at La Jolla, California, during the period August to October, and rainfall during the subsequent rainy season in a portion of southern California. He offers the explanation that low water temperatures at La Jolla indicate that the prevailing winds have been off shore, which results in surface waters being blown out to sea and replaced

by colder waters from below, and that these conditions have the effect of increasing and making for persistence of high pressure off the coast.

On the subject of relationships connected with ice fields, Wiese [68] showed contemporary relationships between Barents Sea ice and temperatures over neighboring land areas some of which gave correlation coefficients of −.80. The influence extends for some distance as indicated by a coefficient of −.58 at Warsaw, and −.60 at Visby. He also worked out a multiple correlation coefficient of .71 between the position of the trough of low pressure off the coast of Norway in January and February and the ice in the previous year in Barents Sea with the rainfall in April and May in central and eastern Russia.

The ice conditions in Barents Sea have been found by Wiese to depend on a number of antecedent conditions extending at least as far south as the equator. The relationships brought out by Wiese suggested to E. H. Smith [69] investigations which showed that a high correlation existed between the amount of field ice in northern waters and the number of icebergs south of Newfoundland, with a coefficient of +.87. He found also a correlation coefficient of −.62 between the number of bergs and the pressure difference between Belle Isle and Ivigtut (Greenland) combined with the deviation of the pressure from normal at Stykkisholm during the period December to March. A coefficient of +.60 was also noted between the number of bergs and the pressure difference Bergen-Stykkisholm during the period October to January inclusive, December being given double weight.

D. Ocean Temperatures.—The effect of ocean temperatures on the climate of coastal regions is recognized by all students of climatology. Just exactly how changes in ocean temperatures are brought about and how such changes affect weather as distinguished from climate is a more complicated question. It has been pointed out that the speed of the trade winds affects the water temperature by blowing away the surface layers and causing some upwelling of the cooler water from below to replace it. Such results are well marked along coasts, on-shore winds being attended by warm surface water along the shore, and off-shore winds causing cold surface waters. The use of ocean temperatures as a basis of relationship with subsequent weather is often coupled with the strength of the winds over the oceans or barometric gradients over the oceans. A work by C. E. P. Brooks [70] is probably the most comprehensive discussion of the subject. In this publication he summarizes the results as follows:

1) The surface temperature of the north Atlantic Ocean between Florida and Valencia has a positive correlation with synchronous pressure over the area, Valencia, Bergen, Berlin, and the Azores, but a negative correlation with pressure at Jacobshavn and Stykkisholm; 2) The pressure at Jabobshavn has a positive correlation with the north east

trade winds 4 months before, this relationship not being due to the influence of the Gulf Stream. 3) The surface temperature of the North Atlantic has a positive correlation with the northeast trade winds 12 months before, this relationship being due to the influence of the Gulf Stream. 4) The surface temperature has a negative correlation with the northeast trade wind 15 to 21 months before. 5) The correlation between the pressure in western Europe and the North Atlantic and the strength of the northeast trade wind 12 to 21 months before is generally small but the coefficients usually have the signs to be expected from relationships 1), 3), and 4); that is, pressure at stations in the area, Valencia, Bergen, Berlin and Azores tends to have a positive correlation with the northeast trade wind 12 months before, and a negative correlation with the northeast trade wind 12 to 21 months before. 6) The surface temperature of the North Atlantic has a positive correlation with the velocity of the southeast trade wind 15 to 21 months before, this relation being due to the influence of the Gulf Stream. 7) Pressure at Valencia, Paris, Berlin and Ponta Delgada has a positive correlation with the velocity of the southeast trade wind 15 to 21 months before; pressure at Jacobshavn, Stykkisholm and Vardo has a negative correlation with the velocity of the southeast trade wind 15 to 21 months before. 8) The surface temperature of the North Atlantic and the pressure at Ponta Delgada have a positive correlation with the Bermuda-Charleston pressure difference 3 to 9 months before and 15 to 18 months before. 9) The surface temperature of the North Atlantic has a positive correlation with the Bermuda-Sydney (Nova Scotia) pressure difference 3 months before; the pressure at Ponta Delgada has a small positive correlation and pressure at Jacobshavn a small negative correlation with the Bermuda-Sydney pressure difference 3 months before. 10) The pressure in western Europe and the North Atlantic (except the Azores) has a negative correlation with the pressure difference 3 months before between the point 50° N., 20° W. and Vestmanno (Iceland). At the Azores the correlation is positive.

Considerable work of a similar nature has been done by others, among them, Meinardus,[71] Hepworth,[72] C. F. Brooks,[73] and Helland-Hansen and Nansen.[74] McEwen [67] has shown a relationship between water temperatures at La Jolla, California, August-October, with precipitation of the following rainy season at 6 stations lying in the region between Los Angeles and San Diego. Predictions have been right as to sign in the great majority of cases but there has been considerable discordance between the magnitude of the actual and the predicted departures. Further, the relationship only holds for a limited region in southern California.

Prior to 1926 minor attention was given to water temperatures except near the surface. The account of the German scientific voyages which was published by Schott,[75] of the Deutsche Seewarte, was discussed by Walker.[76] The latter calls attention to the importance of taking into account aspects of the oceanic circulation other than the surface temperatures. It is pointed out that the distribution of temperature indicates clearly that the surface winds have little or no effect at a depth of 200 m., and inasmuch as the general circulation of the ocean extends to depths of over 3,000 m., it would appear that widespread differences of density, not surface winds, provide the greater part of the motive power for that circulation. Variations in salinity have more effect on density than variations in temperature; in other words, saltiness under prevalent conditions makes the water heavier than coldness. The effects within the topmost 100 or 200 m. are, however, vital in the control of air temperatures. As far as seasonal variations are concerned there is abundant evidence of the importance of a knowledge of ocean conditions. Thus, in the very place where the Swedish expedition vessel "Antarctic" was compressed by the ice in the summer of 1903, an Argentine lightly built vessel moved unhindered in the following spring, meeting no ice. Accordingly, the northern limit of ice in the antarctic may sway backward and forward over 1,500 km. (nearly 1,000 miles). Similar changes occur in the arctic regions where the limit of ice may travel many hundreds of miles in an east-west direction. Thus, 1892 was comparatively free, but 1882 was badly ice bound.

Some seasonal variations may be explained by variations in temperature of a local current that spends some months in passing from the controlling to the controlled station; but variations in activity of the general oceanic circulation will be much more far reaching and important. An abnormally severe season in the antarctic may produce colder currents in the ocean depths, as well as at the surface, and, when this water rises to the surface near the equator, it may affect temperatures there many months afterwards.

From the foregoing discussion any mention of extending the daily forecasts has been omitted. For some years the U. S. Weather Bureau has been issuing on Saturday of each week, "outlooks" of the weather for the coming week. This is about the limit of the extension that can be made with our present knowledge. The usefulness of such "outlooks" has been attested by a considerable demand for them on the part of the industries. In making such predictions, the twice daily maps of the major part of the northern hemisphere have been employed as a basis.

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